A GNEISS DOME IN SOUTHEASTERN VERMONT

by

James B. Thompson, Jr.

A.B. Dartmouth College, 1942

January 20, 1950

SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY IN GEOLOGY

at the

Massachusetts Institute of Technology

Signature of author . . . .
Department of Geology, M.I.T.

Certified by. . . . .
Thesis Supervisor

. . . .
Chairman, Department Committee
on Graduate Students

ABSTRACT

A gneiss-dome east of the Green Mountain anticlinorium in southern Vermont is mantled by metamorphosed strata of probable Cambrian and Ordovician age. The core-rocks are largely gneiss but include mappable horizons of marble, schist, and quartzite, and are believed to be of pre-Cambrian age. The mantle-rocks about the dome are about one-tenth the thickness of the same sequence on the eastern limb of the Green Mountain anticlinorium. Boudinage, stretched pebbles and other phenomena indicate that the thinning about the dome is probably tectonic.

The drag folds on the flanks of the dome show a shear-sense the reverse of that on the limbs of a normal anticline formed by lateral compression, the axial planes of the folds and the planes of secondary slip-cleavage tending to form a flat arch. Analysis of the various structural features indicates that they are inconsistent with anything but upward movement of the core-rock. Possible causes of such upward movement include the density differential between the core-and mantle-rocks while in a plastic state, and a squeezing-upward of the dome rocks as the result of lateral movements at depth.

The zones of regional metamorphism are roughly concentric about the dome with the highest grades (staurolite-kyanite zone) attained near the center. Numerous occurrences of serpentine, soapstone and talc appear to be derived from ultramafic igneous rocks, but clearly antedate the doming and possibly even part of the sedimentation. Small sills and

dikes ranging from quartz diorite to granodiorite in composition are younger than any of the sedimentary rocks, but antedate at least the last stages of deformation and metamorphism. "Snowballed" porphyroblasts and other features indicate the deformation and metamorphism to be essentially contemporaneous.

Small alkalic stocks, including the well-known Mt. Ascutney complex, occur in the same area but appear to be much younger than either the deformation or metamorphism. In the immediate vicinity of the Ascutney stock garnetiferous schists, formed in the regional metamorphism, are recrystallized to a hornfels with relict features of the original schist.

Table of Contents

	Page
INTRODUCTION	1
The Problem	1
Regional Geologic Setting	4
Previous Geological Work	6
Present Investigation	9
Acknowledgements	11
STRATIGRAPHY AND LITHOLOGY	13
General Statement	13
Unconformity Between the Basement and Younger . Rocks	14
Stratified Rocks of the Green Mountain Basement. Complex	18
Gneisses	18
Schists and quartzites	19
Marbles and Ca-Mg silicate rocks	20
Younger Stratified Rocks West of the Green . . Mountain Axis	23
Mendon series	23
Age of the Mendon series	25
Cheshire quartzite	28
Rutland dolomite	28
Champlain Valley sequence above the . . . Rutland dolomite	29
Taconic sequence	29

Page

Younger Stratified Rocks East of the Green Mountain Axis 31

Tyson formation 31

Grahamville formation 33

Pinney Hollow formation 35

Ottauqueechee formation 39

Stowe formation 40

Moretown formation 41

Cram Hill formation and Barnard gneiss . . . 45

Shaw Mountain formation 49

Northfield and Waits River formations . . . 50

Validity and significance of the correlations 54

Stratified Rocks of the Chester Dome 57

Reading gneiss 57

Whitesville marble 59

Cavendish and Gassetts schists 61

Star Hill formation 62

Bull Hill gneiss 63

Grahamville formation 64

Pinney Hollow and younger formations . . . 64

Correlation of the pre-Pinney Hollow formations . . . 68

Unstratified Rocks 70

Ultramafics 70

Granite and pegmatite 74

Veins and minerals pockets 80

Page

Unstratified Rocks (Cont'd)

Greenstone dikes and sills 81

Quartz porphyry dikes and sills 82

Alkalic stocks 83

Trap and felsite dikes 84

STRUCTURE 86

Major Structures 86

Green Mountain anticlinorium 86

Chester dome 90

Minor Structures and Micro-Structures 94

Minor folds 94

Foliation 95

Streaming and mineral lineation 97

Slip-cleavage 99

Rotation and deformation of porphyroblasts . . 101

Pebble deformation 103

Orbicule and quartz-pod deformation 107

Joints 107

Boudinage 108

Relative plasticities of rock-types 109

Analysis of the Deformation 112

Theoretical considerations 112

Rotational features 117

Tensional features 119

Compressional features 121

Page

Analysis of the Deformation (Cont'd)

Pebble deformation 122

Mineral orientation 124

Summary 126

METAMORPHISM 128

Regional Metamorphism 128

Metamorphic zoning. 128

Metamorphism of arenaceous and argillaceous sediments 133

Metamorphism of calcareous sediments 135

Metamorphism of igneous rocks 136

Retrograde metamorphism 137

Movement of material 139

Timing of metamorphism 142

Temperature and pressure conditions 142

Metamorphism at Intrusive Contacts 143

MAJOR PROBLEMS 145

Source of the Taconic Klippe 145

Origin of the Chester dome 146

BIBLIOGRAPHY 150

APPENDIX I, Tabulated . . . . Back pocket

Illustrations

Following page:

Figure 1.	Index map.	9
2.	Stratigraphic column, Wallingford-Clarendon area.	24
3.	Stratigraphic column, Ludlow area . . .	32
4.	Correlation chart	54
5.	Stratigraphic column, Chester dome . .	58
6.	Rotational movements and the strain ellipsoid	118
7.	Boudinage orientation and the strain ellipsoid	120
8.	Minor structural relations, east side of Chester dome	125
9.	Minor structural relations, east limb of Green Mountain anticlinorium	126
10.	Orientation of strain ellipsoid, Green Mountain anticlinorium and Chester dome	127
11.	Metamorphic zoning	128
12.	Composition diagram for argillaceous and arenaceous sediments . . .	129

Plates

Plate I.	Geologic map of southern Vermont	(Back pocket)
Ia.	Geologic cross-section of southern Vermont . .	"
II.	Geologic map of the Ludlow area	"
IIa.	Geologic cross-section of the Ludlow area . .	"
IIb.	Geologic map and cross-section of the Wallingford-Clarendon area	"
III.	Tectonic map of the Ludlow area	"
		Following page
IV.	Unconformity at base of Tyson formation . .	14
V.	Conglomerates of Tyson formation	31
VI.	Pinstripe quartzite of Moretown formation .	42
VII.	Bull Hill gneiss	63
VIII.	Orbicular quartz-diorite, Proctorsville . .	78
IX.	Minor folds, Chester and Ludlow	87
X.	Minor folds, Butternut Hill area	94
XI.	Slip cleavage	99
XII.	Rotated porphyroblasts	101
XIII.	Folded conglomerate, Plymouth	104
XIV.	Deformed pebbles, Plymouth	105
XV.	Boudinage	108

INTRODUCTION

The problem

The field investigations upon which the present study is based were prompted by a general interest in the broader problems of New England geology combined with a more specific interest in the nature of certain bodies of gneiss in the Connecticut Valley region which are characterized by a peculiar domal structure. The gneisses occupying the cores of these domes have been variously interpreted as magmatic, metamorphic or migmatitic in origin, depending upon the inclination of the observer. Despite the variance in interpretation, however, the individual domes are remarkably alike in their stratigraphic, structural and petrologic relations. In the recent literature a magmatic or migmatitic origin has generally been favored. The domes are most numerous in a zone from fifteen to twenty-five miles wide extending from the vicinity of Berlin, New Hampshire, to the north shore of Long Island Sound between New Haven, Connecticut, and the mouth of the Connecticut River. Throughout most of its extent this zone lies just east of the Connecticut Valley. In the area between the Connecticut Valley and the highlands formed by the Green Mountains and Berkshire Hills there is a more scattered distribution of similar structures, not all of which, however, have visible gneissic cores.

In western New Hampshire these domes have an en echelon arrangement along the crest of the Bronson Hill anticline (Billings, 1937). The core gneisses, ranging from granite to granodiorite in composition, constitute the "Oliverian Magma series" of Billings. On the most recent geologic maps of Massachusetts (Emerson, 1917) and Connecticut (Gregory, 1906) the core gneisses are mapped as "Monson granodiorite".

The writer was first introduced to these structures in west-central New Hampshire while an undergraduate at Dartmouth College, and acquired a more detailed knowledge of them during the summer of 1946 while assisting G. E. Moore with field work in southwestern New Hampshire*.

Introduced to the "domes" through the work of Billings and his students, the writer became aware that they constituted one of the key structural and petrologic features of western New England and that any further contribution to the understanding of them would be of great aid in deciphering the complex geology of that area.

Early work by Edward Hitchcock (1861), C. H. Hitchcock (1877), and C. H. Richardson (1902 - 1938) indicated that the gneisses lying between the Green Mountains and the Connecticut River in southeastern Vermont (the Reading gneiss of Richardson) had lithologic and structural

*Doctorate thesis, 1949, Harvard University similarities to the Oliverian gneisses of New Hampshire. As both deformation and regional metamorphism are in general less intense in eastern Vermont than in western New Hampshire, it was felt by the writer that a detailed study of a dome in a less complicated geologic environment than that of western New Hampshire might prove revealing.

The writer's interest in the problem has been shared by John L. Rosenfeld, now a graduate student at Harvard University, since undergraduate days. It was decided, therefore, to do complementary theses and split the burden of the field work, an arrangement that has proved highly stimulating and made possible a much more detailed and extensive field study than could otherwise have been undertaken.

Regional Geologic Setting

The Green Mountains of Vermont are the topographic expression of a major anticlinorium, somewhat complicated by overturning and overthrusting to the west (Plate I). In central and southern Vermont the core of the range is occupied by a crystalline complex of pre-Cambrian age. Ancient crystalline massifs, similar to that of the Green Mountains, also appear in the Berkshires of western Masachusetts, in the Housatonic Highlands of New York and Connecticut and in the Hudson Highlands.

To the west the rocks of the Green Mountain basement complex are overlain with profound unconformity by sedimentary rocks largely of known Cambrian and Ordovician age. In central Vermont a late pre-Cambrian age has been suggested for some of these younger sedimentary rocks (Whittle, 1894; Keith, 1932; Bain, 1938) but recent work, including that of the author, indicates that these beds are more probably Lower Cambrian. In the Champlain Valley, between the Green Mountain anticlinorium and the pre-Cambrian of the Adirondacks, the dominant structural feature is the Middlebury synclinorium (Cady, 1945). The Middlebury synclinorium is overturned toward the west, and bordered by major thrust faults, overthrust in the same direction. South of the latitude of Brandon, Vermont, the synclinorium of the Champlain Valley is complicated by an extensive mass of dominantly argillaceous Cambro-Ordovician sediments which overlies, with apparent structural discordance, the dominantly calcareous and dolomitic Cambro-Ordovician sediments of the Middlebury synclinorium. These rocks, more resistant than the surrounding carbonates, form the Taconic Mountains. The source of this apparently allocthonous mass of rock is one of the major problems of New England geology.

On the east the rocks of the Green Mountain massif are overlain, again with profound unconformity, by two series of metamorphosed sediments and volcanics, separated from each other by a minor unconformity. The upper series, sparsely fossiliferous in some areas, is believed to be Middle Ordovician but may, in part at least, be slightly younger. The lower series has yielded no fossils but is believed, on lithologic and stratigraphic grounds, to range in age from Lower Cambrian possibly to early Ordovician. Throughout most of eastern Vermont these two series form an easterly dipping, homoclinal sequence except where interrupted by the domal structures mentioned previously. Similar domal structures are abundant in western New Hampshire but their relations to those of Vermont are somewhat obscured by the complexities of the Connecticut Valley region.

The dating of the main orogeny in Vermont is uncertain. The youngest well-dated rocks involved are Middle Ordovician. In the Hudson Valley region of New York post-Ordovician folds and faults are truncated by Upper Silurian sediments, dating the Taconic orogeny. In the vicinity of Littleton (New Hampshire), Lake Memphremagog (Quebec), and Bernardston, Massachusetts and elsewhere in New England and southeastern Canada, however, fossiliferous Siluro-Devonian rocks are also highly deformed. According to Billings (1937) most of the crystalline schists of central New Hampshire are to be correlated with the Siluro-Devonian rocks of the Littleton area. Billings believes the main orogeny in New Hampshire to be late Devonian (Acadian). The evidence lies chiefly in the fact that the Quincy granite of eastern Massachusetts, part of a series of unique alkalic intrusives clearly younger than the main orogeny in New Hampshire, Vermont and southern Quebec, is overlain unconformably by the Pennsylvanian sediments of the Narragansett basin (Williams and Billings, 1938). It is probable that both the Taconic and Acadian orogenies affected Vermont although not yet possible to state which was the major event.

Previous Geological Work

The first geologic investigations to make important contributions on the bedrock geology of Vermont were those of Edward Hitchcock and his assistants prior to the publication of his "Geology in Vermont" in 1861. Many of the larger features are delineated correctly on his cross-sections including the anticlinal nature of the Green Mountains and the Chester dome. Although Hitchcock designated most of the metamorphosed strata east of the Green Mountains as "Azoic", the term, as used by him, had no time connotation but referred merely to the non-fossiliferous character of the rocks. Despite his conservative terminology, however, he realized that the Green Mountain gneisses were older than the fossiliferous rocks to the west, and that the conglomerates at the base of the "Talcose schists" east of the mountains were probably equivalent to the conglomerates at the base of the fossiliferous section to the west. The "Clay slates" and "Calciferous mica schist", bordering the "Talcose schists" on the east, were believed by him to be possibly as young as Devonian.

In the twenty years following, interest was largely limited to the relatively unmetamorphosed rocks west of the mountains, shown by Walcott (1888) to be largely of Cambrian age. At about the same time work was begun by the Archean Division of the U. S. Geological Survey in the Green Mountain region of western Massachusetts and southern Vermont. Most of the detailed mapping was in northwestern Massachusetts (Pumpelly, Wolff, and Dale, 1894; Emerson, 1899) but reconnaissance work extended as far north as Rutland, Vermont (Wolff, 1891; Whittle, 1892-1894). The gneisses of the Green Mountain core were identified as pre-Cambrian and the younger strata, both to the west and to the east, as largely Cambrian and Lower Silurian (Ordovician), except for certain strata in the region southeast of Rutland believed by Whittle (1894) to be late pre-Cambrian. Later work in the Green Mountain area by T. N. Dale (1896-1916) added considerably to the detailed knowledge of the area but did not seriously modify the conclusions presented in the earlier work. More recently E. L. Perry (1928), after detailed mapping in the townships of Plymouth and Bridgewater east of the mountains, assigned to the rocks flanking the Green Mountain gneisses on the east ages ranging from late pre-Cambrian to Ordovician, and concluded that structurally the region was essentially an eastward dipping homocline. The first serious modification of the earlier work in the Green Mountain region came with the re-study of a portion of the Taconic-Green Mountain area by Prindle and Knopf (1932). Prindle and Knopf showed that certain of the schists of the Taconic Range, mapped as Ordovician by Pumpelly, Wolff and Dale, were actually Cambrian resting with apparent structural discordance upon strata of Ordovician age. As these Cambrian rocks, together with certain Ordovician rocks of the Taconic Range, represented a distinctly different sedimentary facies from the underlying Cambrian and Ordovician, it was suggested that they had reached their present position by means of a great overthrust with roots somewhere east of the Green Mountains.

A series of papers (1902-1938) by C. H. Richardson dealt chiefly with tracing from the Canadian border to the Massachusetts state line the unconformity within the younger rocks east of the Green Mountains. Richardson considered the rocks west of the line of unconformity (Talcose schists of Hitchcock) to be of Cambrian age and those to the east (Clay slates and Calciferous mica schists of Hitchcock) to be Ordovician on the basis of poorly preserved graptolites in the vicinity of Montpelier. The validity of these fossils, however, has since been questioned by Foyles (1931). Richardson (1928) also recognized the anticlinal nature of the Chester dome and believed that the gneisses in the core were metamorphosed granitic intrusives.

The recent series of investigations in Vermont has now been in progress for about ten years. Published works include those of Currier and Jahns (1941), Hawkes (1941), Doll (1944), Cady (1945), and Kaiser (1945). The bulk of the work, however, is either still in progress or awaiting publication (Index Map, Figure 1.)

Present investigation:

The writer spent approximately nine months in the field during the summers of 1947, 1948 and 1949, covering an area of about three hundred square miles. The area of detailed mapping is on the eastern slopes of the Green Mountains in the vicinity of Ludlow and Chester, Vermont, and includes, geologically, the northern half of the Chester dome and a section of the eastern limb of the Green Mountain anticlinorium. The topographic maps covering the area are the Ludlow quadrangle (entire sheet), Clarement quadrangle (part of western third) and Wallingford quadrangle (part of eastern

Figure 1. Index Map

Key:

1. Ludlow area.
2. Wallingford-Clarendon area.
3. Saxtons River area (J. L. Rosenfeld, work in progress)
4. Woodstock area (P. H. Chang, work in progress)
5. Hanover area (J. B. Lyons, work in progress)
6. Claremont area (C. A. Chapman, 1942)
7. Bellows Falls area (Kruger, 1946)
8. Keene-Brattleboro area (Moore, 1949)
9. Plymouth-Rochester area (Hawkes, 1941)
10. Rochester area, northern part (P. H. Osberg, work in progress)
11. Strafford area (Doll, 1944)
12. Barre area (Currier and Jahns, 1941)
13. Lincoln Mtn. area (W. M. Cady, work in progress)
14. Montpelier area (W. M. Cady, work in progress)
15. Brome County area (Fairbairn, 1932)
16. West-central Vermont (Cady, 1945)
17. Castleton area (Fowler, 1949)
18. Northern Taconic area (Kaiser, 1945)
19. Taconic quadrangle (Prindle and Knopf, 1932)
20. Dutchess County area (Balk and Barth, 1936)
21. Shelburne Falls area (Balk, 1946)
22. Wilmington area (J. W. Skehan, work in progress)

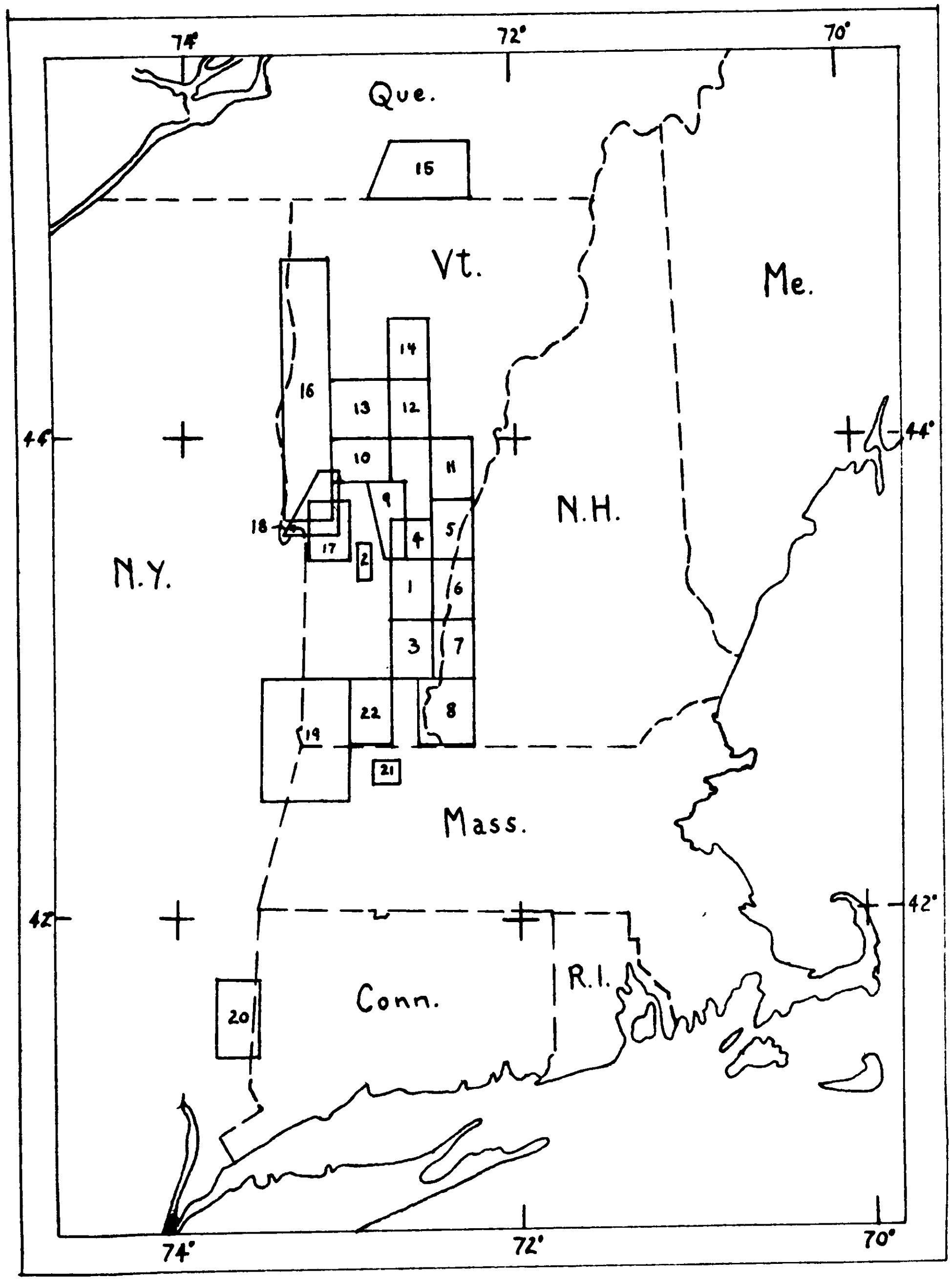


Figure 1. Index Map

(Key on opposite page.)

third) all on a scale of 1:62,500. In addition to the work on the east side of the Green Mountains a series of traverses was made along the western front of the Green Mountains between North Clarendon, in the southwestern part of the Rutland quadrangle, and South Wallingford, in the northwestern part of the Wallingford quadrangle.

The area mapped by Rosenfeld includes most of the Saxtons River quadrangle to the south and part of the Londonderry quadrangle to the southwest. Rosenfeld's mapping covered the southern part of the Chester dome as well as the adjacent section of the eastern limb of the Green Mountain anticlinorium. The work of P. H. Chang in the Woodstock quadrangle and eastern part of the Rutland quadrangle has provided continuity to the north. Concurrent geologic mapping has also been carried on in the Claremont quadrangle, to the east, by C. A. Chapman, and in the Hanover quadrangle, to the northeast, by J. B. Lyons. The Bellows Falls quadrangle, to the southeast, has been mapped recently by F. C. Kruger (1946).

The topographic relief varies from about five hundred feet in the Connecticut Valley region to about two thousand feet in the Green Mountain region where the higher summits generally exceed three thousand feet in elevation. The outcrop in stream beds and on ridges and hilltops is in general good although many of the main river valleys are choked with till and fluvio-glacial deposits.

The spacing of traverses varied from half a mile to a mile or more depending upon the complexity of the local structures. Outcrops were located largely by means of compass and aneroid supplemented by pacing when necessitated by flat terrain. The laboratory phase of the investigation has consisted largely of routine microscopic study of thin sections and mineral grains.

Acknowledgments

The writer wishes to state that the field work would probably have been impossible without the able and enthusiastic assistance of fellow students during each of the field seasons. The writer wishes to thank, in particular, Ping Hsi Chang and John Gieling of Harvard University for assistance during the summer of 1947; Robert J. Bean of Harvard and K. N. Das of M.I.T. for assistance during the summer of 1948; and James W. Skehan and Arden Albee of Harvard and Mark C. Wittels of M.I.T. for assistance during the 1949 season. To others who aided the writer on weekend expeditions and at other times proportionate thanks are extended.

The writer also wishes to state his indebtedness to: Professor Harold W. Fairbairn for encouragement and advice as thesis advisor and for first awakening in him an active interest in the complex problems of petrology; Professor Martin J. Buerger for introducing him to the science of crystal chemistry and its possible applications to these same problems; Professor Marland P. Billings of Harvard University for first acquainting him with the problems of New England geology and for stimulating field consultations; and John L. Rosenfeld, Ping Hsi Chang, Wallace M. Cady, Carleton A. Chapman, John B. Lyons, Philip H. Osberg, James W. Skehan and other fellow workers in Vermont for their friendly co-operation in thrashing out mutual problems in the field.

LITHOLOGY AND STRATIGRAPHY

General Statement

The choice of terms is something of a problem to the geologist who wishes to describe crystalline rocks without bringing in unwanted genetic implications. "Sedimentary", "igneous", "intrusive", "extrusive", pyroclastic" and "metasomatic" all have too many genetic strings attached, as do the prefixes "ortho-" and "para-". "Metamorphic" is much too general. Faced with this dilemma the writer has resorted to terms regarded by many, no doubt, as antiquated, but which do have the advantage of enabling one to classify the rocks on a descriptive basis without overly committing one's self as to their origin.

For the preliminary breakdown the writer has chosen the terms "stratified" and unstratified" following Edward Hitchcock (1861) and other nineteenth century geologists. By "stratified" the writer means any rock exhibiting pronounced compositional banding, regardless of its origin. The stratified rocks of the Ludlow area include rocks believed to have been originally sediments, pyroclastics, igneous extrusives and possibly a few banded intrusives, modified to a varying degree by later metamorphic and metasomatic processes. The unstratified rocks occur, for the most part, as cross-cutting masses and are believed to be largely of magmatic or anatectic origin.

The thicknesses given in the following pages for the various stratigraphic units are the present thicknesses measured normal to the dip without correction for tectonic thickening or thinning. The writer believes, however, that in the Green Mountain area the thicknesses given are probably not far from the original stratigraphic thicknesses of the various units.

The lithologic descriptions in the following text have been limited largely to the megascopic features by means of which the formations were differentiated in the field. For the quantitative mineralogy of the more important rock types the reader is referred to the tabulated modes (Appendix I.)

Wherever possible, suitable formation names already in the literature have been retained, in some instances with minor revisions. The stratigraphic nomenclature for Vermont, however, is in a state of flux at the present time and it is likely that many of the names used here will be superseded when the various geologists now working in Vermont have agreed upon a revised nomenclature.

Unconformity between the basement complex and younger rocks.

Although probably the most important break in Vermont stratigraphy, the unconformity between the basement complex of the Green Mountain massif and the younger metamorphics has not always been recognized. Its existence has been denied by some recent investigators (Foyles, 1928-33; Hawkes, 1941)

Plate IV.

Unconformity at base of Tyson formation

The locality is about three quarters of a mile west of Ludlow village. The rocks in the foreground are contorted gneisses and sheared pegmatites (light patches) of the Green Mountain basement rocks. The basal units of the Tyson formation, immediately overlying, include arkosic conglomerate, mica schist and thin-bedded impure quartzite.

A

B

and even those who have admitted its existence have not always agreed upon its exact location. The reason for this surprising state of affairs concerning such a profound stratigraphic break is in part structural and in part metamorphic.

The structural difficulties lie largely in the tendency of the later orogenic movements to rotate the planar features in the older rocks into sub-parallelism with those of the younger, and the metamorphic difficulties in the fact that the arkoses and conglomerates forming the base of the younger section, when metamorphosed, resemble very strongly much of the banded granitic gneiss of the basement complex.

Because of these major difficulties it is only in a few localities where later events have not almost completely obliterated the evidence that the unconformable relations are clear-cut. Until one of these localities has been seen it is possible to cross the unconformity many times and never suspect its existence. Once its existence has been demonstrated however, it is not too difficult to trace it from the clear-cut localities into regions where the relationships are more subtle.

The evidence for a profound angular unconformity in the Green Mountain region is now believed to be conclusive inasmuch as the angular relationships are actually visible at several localities. (Plate IV) Furthermore, the older rocks are cut by bodies of granite, coarse pegmatite, and by coarse quartz-tourmaline veins which do not appear in the younger rocks in the vicinity, but fragments of which do appear as pebbles in the basal conglomerates of the younger series. A particularly useful criterion is the conspicuous "blue" quartz characteristic of much of the pegmatite and vein material in the older rocks, but appearing only as pebbles in the younger. The basement rocks are also of distinctly higher metamorphic grade, in the Green Mountain area, and are more coarsely crystalline, showing fewer traces of original sedimentary features. The dolomites of the basement complex, for example, contain coarse crystals of diopside, actinolite, phlogopite and other Ca-Mg silicates whereas those of the younger rocks, immediately overlying contain only a few minute scales of phlogopite. Large-scale retrograde features, probably dating from the later, lower grade metamorphism, are also common in the older rocks and are not found in the younger. These include pseudomorphs of actinolite after diopside, talc after actinolite, chlorite after garnet and saussurite after plagioclase. The last, because it is widespread and easy to spot, is a useful field criterion. Because of the evidence outlined above, the unconformity is believed to represent not only a major time gap but also a period of deformation accompanied by regional metamorphism and emplacement of masses of granite and pegmatite. As the overlying formations include beds of known Lower Cambrian age, the pre-Cambrian age of the basement complex is convincingly demonstrated.

In the Chester dome, however, the Paleozoic deformation and metamorphism have been much more intense than in the Green Mountain region, and the obliteration of the unconformity is consequently much more complete. In fact, the writer must admit that nowhere in the Chester dome area has he seen any convincing field evidence for such a break. It has, therefore, been necessary to rely wholly upon indirect stratigraphic,structural and lithologic evidence in assigning to the unconformity in the dome area its present provisional location.

Stratified Rocks of the Green Mountain Basement Complex

Gneisses

Gneisses of various sorts are definitely the dominant rock types of the basement complex in the Green Mountains, so much so, in fact, that the entire complex has often been referred to as the "Green Mountain gneiss". Most common are biotite gneisses containing both microcline and sodic plagioclase in varying proportions. Locally these grade into highly muscovitic or hornblendic gneisses. Both the muscovitic and hornblendic types are generally lower in potash feldspar than the more abundant biotitic types, the potash apparently taken up by the micas in the muscovitic gneisses and simply less abundant in the hornblendic types. Most of the gneisses are coarse-grained, well-foliated and well-banded. The banding is evident chiefly in the kinds and proportions of dark minerals present.

The most conspicuous microscopic feature is the intense alteration of the plagioclase feldspars, and the ferromagnesian minerals. The plagioclase, particularly in the more hornblendic gneisses, is generally filled with small rods of epidote and scattered flakes of sericite, somewhat coarser than a typical saussurite and not uncommonly showing a zonal arrangement with the more epidotic zones toward the center of the grain. This alteration is usually apparent in hand specimen from the green color of the epidote or, if the epidote is not strongly colored, by the chalky appearance of the feldspars on a fresh surface. The potash feldspars, generally microcline, contain abundant coarse sericite usually oriented parallel to visible cleavage traces. The biotites are commonly altered to chlorite, and the hornblendes to chlorite, epidote and rusty-weathering carbonate. These alteration products are conspicuously coarser-grained and the constituent minerals more easily identified than the typical products of the deuteric alterations of igneous rocks.

Schists and quartzites

Interbedded schists and quartzites rank second in abundance to the gneisses in the Green Mountain basement complex and have been roughly differentiated from the latter on the map.

The quartzites are most abundant in the belt of schists extending north-northeasterly from Ludlow Mountain. These quartzites are very pure and are recrystallized to the extent that they have an almost glassy appearance on a fresh fracture. The individual quartzite beds vary in thickness from a few inches to one hundred feet or more. Interbedded with the quartzites are highly quartzose, generally graphitic muscovite schists. Some of the schists contain small amounts of chlorite, biotite and small garnets. Others contain small augen of plagioclase (albite) varying from a sixteenth of an inch to nearly a quarter inch in long diameter. In some of the non-graphitic quartz-muscovite schists, as well as in some of the purer quartzites, octahedra of magnetite up to a quarter inch in diameter are not uncommon.

The schists and quartzites of the Terrible Mountain area show significant differences from those of the Ludlow Mountain area. The quartzites are similar, but thinner and less abundant and the schists are more highly chloritic and garnetiferous with fewer graphitic beds. There are also several horizons of amphibolite and hornblende gneiss interstratified with the schists of Terrible Mountain and none in the schists of the Ludlow Mountain area. One of these bands of hornblende gneiss, on the western slopes of Terrible Mountain, in the Wallingford quadrangle, is remarkable in the occurrence of large almandite garnets, some of which attain a diameter or two inches or more. Locally these garnets are partially or completely altered to chlorite. Except for the alteration of the garnets the rock is very similar, at least in hand specimen, to the well-known garnet-rock from Gore Mountain near North Creek, N. Y.

Marbles and Ca-Mg silicate rocks

Marbles and Ca-Mg silicate rocks are of common occurrence in the band of gneisses bordering the Ludlow Mountain - Tiny Mountain schists and quartzites on the east. The rock types include dolomite marble, calcite marble, actinolite-diopside rock, actinolite-phlogopite rock and talc schist, interstratified with quartzites, garnetiferous mica schists, and a variety of gneisses. The calcitic marbles are coarse grained and suggest, by their close association with the Ca-Mg silicate rocks, that the calcite is the result of "de-dolomitization". A remarkable feature of many of the Ca-Mg silicate rocks of the basement complex is their coarseness of grain, the diopside, actinolite and phlogopite crystals averaging, in some occurrences, between one and two inches in diameter. Epidote and Sphene are common, the epidote at one locality west of Grahamville occurring as an overgrowth on grains of allanite. The talc schists are somewhat unusual in that the talc has a fibrous habit suggesting that it may be pseudomorphous after an amphibole, possibly tremolite, and fibrous aggregates of actinolite showing the characteristic parting and crystal habit of diopside are also common. Similar occurrences of marble and Ca-Mg silicate rocks are common on the basement rocks of the Wallingford quadrangle. Some of these have been described by Whittle (1894), Eggleston (1918), Dale (1915) and Perry (1928), and additional occurrences have been found by the writer. In addition to the types listed above, actinolitic and calcitic gneisses are fairly common.

At the "Devil's Den" near the height of land on the road between Danby and Weston in the Wallingford quadrangle, a thick bed of buff-colored dolomite is apparently overlain by quartzites and banded muscovitic gneisses and underlain by a coarse biotite-muscovite-albite schist. Both the schists and the quartzites and gneisses contain coarse quartz-tourmaline veins and pods of pegmatite. Although the dolomite, in its purity and relative lack of secondary silicates, resembles some of the younger dolomites more than those of the basement complex, the lithologic sequence is inconsistent with that in the younger rocks on either side of the range, and the grade of metamorphism shown by the associated schists and gneisses is higher than that shown by any of the known younger rocks of the Green Mountain area. The sequence does strongly resemble, however, although in inverted order, the Reading gneiss, Whitesville marble and Cavendish schist of the Chester dome.

Stratified Rocks West of the Green Mountain Axis

Mendon Series

The basal unit of the Mendon series in the Wallingford-Clarendon area on the west flank of the Green Mountains consists of conglomerate, greywacke, arkose and impure quartzite with an average total thickness of about 600 feet. The conglomeratic zones are lenticular in nature and are most abundant near the base of the formation. The pebbles reach a maximum diameter of eight to ten inches where least deformed, and are composed of quartzite, vein quartz, pegmatite, and gneiss. Quartzite pebbles are by far the most abundant, particularly in the coarser phases of the conglomerate. All of the rock types found in the pebbles are characteristic of the underlying Mount Holly series. Pebbles of blue, opalescent quartz identical with that in most of the quartz-veins and pegmatites cutting the Mount Holly series are abundant. The matrix of the conglomerates is generally a quartz-muscovite schist. The most conspicuous accessory is magnetite, occurring as octahedral prophyroblasts up to a quarter inch in diameter. Some beds may carry as much as two or three percent magnetite. The quartzites are generally muscovitic and schistose and grade, with increasing feldspar content, into arkose or greywacke. The greywackes contain sub-angular pebbles of blue quartz and alkali feldspar varying from an eighth to a half inch in diameter, in a matrix of quartz-sericite-biotite (and / or chlorite) schist containing rhombs of calcite or ankerite. The carbonate content is variable and a few lenses of pure dolomite and calcite marble, generally not more than four or five feet thick, have been found. In the gorge at East Clarendon a bed of dolomite breccia about ten feet thick occurs about one hundred feet below the top of the formation. The quartzites associated with the dolomites and dolomitic arkoses in the upper part of the formation are somewhat purer than those below, and on Bear Mountain in Wallingford a bed of pure white quartzite about six feet thick was observed at this horizon.

The upper unit of the Mendon series in the Wallingford-Clarendon area is largely dark, graphitic phyllite. In the few non-graphitic zones the phyllite has a pale green color. In its upper portions the phyllite becomes sandy and contains interbeds of pure white quartzite from five to twenty feet thick identical lithologically with the overlying Cheshire quartzite. The phyllites and associated quartzites are between 300 and 400 feet thick.

The basal arkose and conglomerate of the Mendon series has been named Nickwacket greywacke by Keith (1932) after exposures on Mt. Nickwacket east of Brandon. The Nickwacket greywacke is probably continuous with the Pinnacle arkose (Clark, 1936) of southern Quebec and the Ripton conglomerate of Dale (1910) is probably a conglomerate at or near the base of the Nickwacket. The Dalton formation (Emerson, 1917) of western Massachusetts is similar in both lithology

Figure 2.

Stratigraphic Column for Wallingford-Clarendon Area

Formation:	Thickness:	Lithology:	Age:
Rutland dolomite	1,000' +	Grey to buff-colored dolomite	Lower Cambrian
Cheshire quartzite	400' ±	Thick-bedded white quartzite	Lower Cambrian
Mendon series (upper part)	300 - 400'	Graphitic phyllite, thin-bedded impure quartzite, massive white quartzite	Probably Lower Cambrian
Mendon series	600' ±	Conglomerate, greywacke, arkose, impure quartzite, with lenses of calcite and dolomite marble near top	Probably Lower Cambrian
------ Angular Unconformity ------			
Green Mountain Complex	---	Gneiss, schist, quartzite, marble, Ca-Mg silicate rock, granite, pegmatite	Pre-Cambrian

and stratigraphic position to the conglomerate and arkose of the Mendon series. The phyllites of the upper part of the Mendon series are believed to be equivalent to the Moosalamoo phyllite of the Brandon area and the West Sutton slate of southern Quebec. In Massachusetts argillaceous rocks associated with the Cheshire quartzite and Dalton formation have been mapped as either Hoosac schist (on the west slope of Hoosac Mountain) or as Berkshire schist. In the Brandon area the Forestdale marble of Keith (1932) separates the Nickwacket greywacke and Moosalamoo phyllite. It is probably equivalent to the White Brook dolomite of Quebec and to the carbonate beds near the top of the conglomerate arkose zone at Mendon (Whittle, 1894) and East Clarendon. No carbonate lenses were observed by the writer, however, south of the gorge at East Clarendon.

Age of the Mendon Series

The only place where fossils have been found in rocks possibly equivalent to the Mendon series is on Clarksburg Mountain near Williamstown, Massachusetts just south of the Vermont-Massachusetts state line. Here Walcott (1888) found trilobite fragments, probably *Olenellus*, in a quartzite containing pebbles of blue quartz at a point about one hundred feet above the unconformity with the gneisses of the Green Mountains. Prindle and Knopf (1932 p 271-2) describe the stratigraphy as follows:

"In several places the basal Cambrian beds are conglomeratic with pebbles stretched along a foliation plane parallel to the bedding. The basal beds, whether schistose or conglomeratic, grade upward into schistose feldspathic quartzites with interbedded conglomerates. About 100 feet above the base of the formation Walcott discovered fragments of Olenellus. Over the Olenellus-bearing beds are thin to thicker-bedded pure white quartzite that show sharp crumpling of the individual beds into acute folds. Above these quartzites are micaceous phyllites in places carbonaceous". ---- "The phyllites on Clarksburg Mountain are overlain by pure, heavy-bedded quartzite, which forms the lowest slope of the mountain plunging southward under the Rutland dolomite of the valley".

---Although the entire sequence is mapped as "Cheshire quartzite" by Emerson (1917), and by Prindle and Knopf (1932), it is apparent that it resembles more closely the Cheshire quartzite plus Mendon series of the Wallinford-Clarendon area. As nearly as can be determined the thicknesses of the various units in the two areas are approximately equivalent. Despite this fact, however, Whittle (1894) suggested a pre-Cambrian age for the Mendon series apparently believing the entire sequence on Clarksburg Mountain to be the equivalent of the quartzite overlying the Mendon series.

Whittle gives as additional evidence for an unconformity between the Mendon series and the Cheshire quartzite, the more intense minor deformation in the Mendon series, and the fact that overturned dips were common in the Mendon series and not found by him in the overlying rocks. It is true that the Cheshire quartzite shows less minor deformation than the phyllites and schistose arkoses of the Mendon series, but it is also true that it is a much more competent rock. The dolomites above the Cheshire show minor deformation fully as intense as any in the Mendon series. The overturned dips in the Mendon series are of doubtful stratigraphic significance inasmuch as the Cheshire quartzite is also overturned south of Clarendon Gorge. Although the writer has seen no evidence for a discordance between the Cheshire quartzite and the Mendon series. Foye (1919) and Keith (1932) describe evidence of such a discordance in a section east of Lake Dunmore near Brandon. At this point the Forestdale marble is overlain by a conglomerate interpreted by Foye and Keith as basal Cheshire. If so, this would indicate the existence of an unconformity between, as the Moosalamoo phyllite would be absent. The writer doubts, however, that the conglomerate is basal Cheshire. The base of the formation is definitely not conglomeratic in the Wallingford-Clarendon area. It is also to be noted that the carbonate lenses believed to correspond to the Forestdale marble are, both at Mendon (Whittle, 1894, pp. 410-411) and east of Clarendon, <u>near</u> but not <u>at</u> the top of the conglomerate-arkose zone.

The writer is of the opinion that the age of the Mendon series is Lower Cambrian.

Cheshire quartzite

The Cheshire quartzite, named from the type locality in Cheshire Massachusetts, is, in the Wallingford-Clarendon area, a pure, massive, fine-grained quartzite about 400 feet thick. At some localities in Massachusetts phyllite, arkose and conglomerate have been included but are probably equivalent to rocks mapped as Mendon series in Vermont. Keith gives the thickness of the formation as eight hundred feet at Wallingford, but from his lithologic description it is likely that he included a large part of the Mendon series as defined by Whittle. Lower Cambrian fossils have been found in quartzites assigned to the Cheshire at several points in western Vermont (Walcott, 1888). It is possible that at Clarksburg Mountain the quartzites may prove to belong to the Mendon series, but this is not critical in dating the Cheshire inasmuch as the overlying dolomites are also Lower Cambrian. The Cheshire quartzite is equivalent to the Gilman quartzite (Clark, 1936) of southern Quebec.

Rutland dolomite

Only the basal beds of the Rutland dolomite were studied by the writer. At Wallingford and Clarendon these consist of alternating beds of buff and dark grey dolomite containing minute scales of phlogopite. The dolomite contains Lower Cambrian fossils (Wolff, 1891) in the vicinity of Rutland. The Dunham dolomite of southern Quebec and northwestern Vermont is probably equivalent (Cady, 1945) to the lower portion of the Rutland dolomite and the Rutland, in turn, equivalent to the lower portion of the Stockbridge limestone of Massachusetts. The contact with the Cheshire quartzite has not been seen by the writer. According to Cady (1945) it is a transition zone about 50 feet thick consisting of interbedded quartzite, dolomite, and dolomitic quartzite.

Champlain valley sequence above the Rutland dolomite

The strata overlying the equivalents of the Rutland dolomite in the Champlain valley have been described in detail by Cady (1945), and range in age from late Cambrian to Middle Ordovician. Carbonates predominate and are chiefly dolomites in the Cambrian, and limestones with some dolomites in the Ordovician. There is a major stratigraphic break in the Middle Ordovician and the Hortonville shale, of Trenton age, overlies the Cambrian and early Ordovician carbonates with angular unconformity. Some of the carbonates unconformably beneath the Hortonville shale are as young as Black River and even early Trenton, indicating that the break lies within the Trenton.

Taconic sequence

The rocks of the Taconic Range rest with structural discordance upon those of the Champlain valley, forming a gigantic klippe covering a large area in western Vermont, eastern

New York State, and western Massachusetts. The rocks of the Taconic sequence, like those of the Champlain valley sequence, range in age from Lower Cambrian to Middle Ordovician with an unconformity in the Middle Ordovician. Lithologically, however, the Taconic rocks are quite different from the Champlain Valley rocks, being dominantly argillaceous in contrast to the carbonates of the valley sequence. The Taconic sequence rests with pseudo-conformability upon the Hortonville shale, the youngest formation of the Champlain Valley sequence. This relationship has caused considerable confusion inasmuch as the unconformity at the base of the Hortonville is the more conspicuous feature.

Younger Stratified Rocks East of the Green Mountain Axis

Tyson formation

The Tyson formation, named for the excellent exposures of the conglomeratic phase of the formation northwest of Tyson Village, is the basal unit of the younger stratified series on the eastern limb of the Green Mountain anticlinorium. Lithologically it is almost identical with the basal units of the Mendon series to the west. The principal lithologic types are conglomerate, schistose greywacke, micaceous quartzite and arkose. Near the top of the formation beds of pure quartzite, quartz-rich sericite-chlorite schist and calcite marble also occur. The formation varies from zero to almost 600 feet in thickness, the greatest thicknesses occurring on Dry Hill in Plymouth, and on the eastern slopes of Ludlow Mountain. Midway between these two localities, however, the formation is entirely absent and the schists of the Grahamville formation rest directly upon the older gneisses. West of Dry Hill a similar situation occurs in that only the upper, calcite marble member of the Tyson formation is present.

The conglomerates (Plate V) are best developed on Dry Hill. The pebbles are predominantly quartzite with minor amounts of vein-quartz, gneiss, granite and pegmatite. The quartzite pebbles contain black tourmaline in isolated crystals

Plate V.

Conglomerates of Tyson formation

The locality is at an elevation of about 1500' on the eastern side of Dry Hill in Plymouth. The pebbles are chiefly quartzite with a few of gneiss. The large quartzite boulder above the hammer handle in the upper photograph measures about 9 x 27" in cross-section. The elongation is down the dip (away from the camera). The axial ratios of the pebbles average about 1 : 3 : 9 in this area.

A

B

and in small veins. In several instances tourmaline crystals were observed to be truncated at the surface of the pebble. The occurrence of tourmaline is identical with that in the underlying schists and quartzites of the Ludlow Mountain area. At one locality on the eastern slopes of Ludlow Mountain a quartz conglomerate of the Tyson formation may be seen resting directly upon these older quartzites. Most of the pebbles are flattened in the plane of foliation and elongate in the direction of the dip. The largest pebbles observed measured ten by thirty inches in cross section and were at least four feet long. The matrix of the conglomerate is generally a quartz sericite schist containing magnetite in perfect octahedra about 1/8 inch in diameter.

The schistose greywackes make up the bulk of the formation in most areas. These consist of rounded to sub-angular pebbles of quartz, microcline, orthoclase, and sodic plagioclase in a matrix of quartz - sericite - biotite schist. The minor constituents include chlorite, carbonates, epidote and magnetite. The pebbles average between 1/4 and 1/8 inch in diameter and commonly show a regular decrease in size toward the top of a bed. Some of the more pebbly greywackes and arkoses resemble augen gneisses, particularly if the pebbles are somewhat crushed. Many of the plagioclase pebbles have an altered core containing sericite and epidote, surrounded by a mantle of fresher material. It is possible that these represent secondary overgrowths on the original detritital

Figure 3.

Stratigraphic Column for Ludlow Area

Formation	Thickness	Lithology	Age
Waits River formation	5,000' +	Blue-grey siliceous limestone, garnetiferous graphitic phyllite	Middle Ordovician, probably Trenton
Northfield formation	500' - 700'	Garnetiferous graphitic phyllite, minor beds of blue-grey siliceous limestone	" "
Shaw Mtn. formation	0 - 500'	Quartz conglomerate quartzite, quartz-sericite-garnet schist, greenstone schist, amphibolite	" "
--------- Unconformity --------			
Barnard gneiss	3,500' - 6,200'	Biotite gneiss, biotite-hornblende gneiss, hornblende gneiss, garnetiferous hornblende gneiss, amphibolite	Cambrian or early Ordovician
Cram Hill formation	3,500' - 6,200'	Graphitic and non-graphitic quartz-sericite-garnet schist , graphitic quartzite , greenstone schist , amphibolite	" "
Moretown formation	3,500' - 3,700'	"Pinstripe" quartzite, greenstone schist fasciculitic hornblende schist	" "

Formation	Thickness	Lithology	Age
Whetstone Hill member, Moretown formation	1,600'	Irregularly graphitic quartz-sericite-biotite-garnet-plagioclase schist, Manganiferous ironstone, "pinstripe" quartzite, greenstone schist, fasciculitic hornblende schist	Cambrian or early Ordovician
Stowe formation	600 - 900'	Pale green quartz-sericite-chlorite-schist, greenstone schist	Probably Lower Cambrian
Ottauqueechee formation	500 - 1,400'	Graphitic schist and quartzite, quartz-sericite-chlorite schist, greenstone schist	" "
Pinney Hollow formation	1,600 - 3,000	Pale green quartz-sericite-chlorite schist, greenstone schist, graphitic schist, albitic quartz-sericite-chlorite schist	" "
Grahamville formation	800 - 2,700'	Generally graphitic quartz-sericite-albite-biotite schist and quartz-sericite-albite-chlorite schist	" "
Plymouth member, Grahamville formation	0 - 1,000'	Massive white quartzite, thin-bedded schistose quartzite, dolomitic quartzite, dolomite, dolomite breccia	" "

Formation	Thickness	Lithology	Age
Tyson formation	0 - 600'	Conglomerate, greywacke, arkose, thin-bedded schistose quartzite, quartz-sericite-albite-chlorite schist, lenses of calcite marble in upper part	Probably Lower Cambrian

----------Angular Unconformity-----------

Green Mountain Complex		Gneiss, schist, quartzite, marble, Ca-Mg silicate rock, amphibolite, granite, pegmatite	Pre-Cambrian

grains. The carbonate, commonly occurring as small rhombic grains, may be either calcite or ankerite. Chlorite is much less abundant than biotite and has not been observed in rocks containing a potash feldspar.

Following Hitchcock (1861) the Tyson formation was correlated with the basal conglomerates and greywackes of the Mendon series to the west, and is, therefore, of probable lower Cambrian age.

Grahamville formation

The albitic schists overlying the Tyson formation in the Ludlow area have been named the Grahamville formation. Probably the best exposures of these albitic schists are in the railroad cuts on the sharp bend in the tracks about a half mile south of Grahamville Village in the north-central part of Ludlow. The Grahamville formation is equivalent to the upper part of the "Older Cambrian group" of Perry (1928) and to much of the Plymouth Union series of Hawkes (1941), but inasmuch as Hawkes included rocks now known to belong to the basement complex as well as equivalent to the Tyson formation, it has been deemed advisable to introduce a new name. The quartzite-dolomite horizon in the upper part of the formation has been mapped separately as the Plymouth member in accordance with the frequent references to "Plymouth marble" in the literature. The Plymouth member is equivalent to the quartzite and dolomite of Perry's Older Cambrian series, but more recent work by P. H. Chang (personal communication) and the writer has shown that the three quartzite-dolomite horizons of Perry are actually one with structural duplication.

The thickness of the Grahamville formation is about 2,700 feet at Ludlow and about 800 feet in the southern part of Andover. The Plymouth member is about 1000 feet thick in Plymouth and Ludlow but appears to be non-existent south of the Ludlow-Andover line. The schists constituting the bulk of the formation are essentially quartz-sericite-biotite-albite schists with minor amounts of chlorite and magnetite. Minute needles of tourmaline, barely visible to the naked eye, are not uncommon and much of the schist is graphitic. The albite occurs as untwinned augen, the smallest just visible with a hand lens and the largest measuring about 3 x 5 mm. in cross section. The schists immediately below the Plymouth member are relatively rich in quartz and commonly contain small amounts of carbonate.

The Plymouth member includes quartzite, dolomitic quartzite, dolomite breccia, and minor beds of albite schist. The quartzites and dolomitic quartzites chiefly underlie the main dolomite horizon. Most of the quartzites are slightly feldspathic and schistose with purer quartzites occurring in large lenses. The pure quartzites are white or buff-colored and resemble the Cheshire quartzites of the Wallingford-Clarendon area. The dolomites are grey or buff-colored and occur in beds generally less than 50 feet thick, separated by thin-bedded quartzites and albite schists. Most are siliceous and contain small flakes of phlogopite. A dolomite breccia about 75 feet thick on the east side of Lake Amherst was formerly quarried for ornamental stone. It contains fragments of buff-colored dolomite in a matrix of dark grey dolomite. According to Dale (1915) the dark color is caused by disseminated flakes of graphite.

The Grahamville formation is probably equivalent to the Hoosac schist of Massachusetts and the lower part of the Sutton schists of southern Quebec. The lower albite schists are correlated with the upper, phyllitic portion of the Mendon series, and the quartzites and dolomites of the Plymouth member with the Cheshire quartzite and the lower part of the Rutland dolomite respectively.

Pinney Hollow formation

The Pinney Hollow formation (Perry, 1928) is a pale green quartz-sericite-chlorite schist interstratified in its upper part with greenstone schists. There are also, just above the middle of the formation, a few thin beds of graphitic phyllite. The thickness of the formation varies from 3,000 feet in Ludlow in Plymouth to 1,600 feet in the southern part of Andover. The portion above the horizon of the graphitic phyllites, largely greenstone schists, is about 1,300 feet thick in Ludlow and about 800 feet thick in southern Andover.

In addition to quartz, sericite and chlorite the Pinney Hollow schists commonly contain magnetite, occurring as nearly perfect octahedra one or two mm. in diameter, and tiny needles of tourmaline. The magnetite content may run as high as three or four percent. Some of the schists are albitic, the albite occurring as small phorphyroblasts as in the underlying Grahamville schists. In regions of higher grade metamorphism, as in the southern part of Plymouth, the schists are garnetiferous and biotitic, the garnet (almandite) occurring as dodecahedra about 5 mm. in diameter and the biotite as glossy black, rhombic plates set transverse to the foliation. In this same area the base of the formation also contains schists rich in chloritiod. According to Perry (1928, p. 26) some of the schists on Blueberry Mountain in Plymouth contain staurolite and kyanite but the writer was unable to confirm this observation.

Quartz is the chief constituent of most of the Pinney Hollow schists. In addition to the fine-grained quartz of the matrix the schists are characterized by odd paddle-shaped masses of "vein" quartz, measuring from a quarter to a half inch in thickness, three to six inches in width and as much as two or three feet in length. These quartz lenses are oriented parallel to the foliation with their long axes approximately down the dip.

The greasy feel of the sericite led early geologists to believe that the mineral was talc, hence the name "talcose schist". The sericite is definitely an aluminous mica, however, and not talc.

The quartz-sericite-chlorite schist of the Pinney Hollow are believed to be metamorphosed ferruginous shales. The high magnetite content of some of the schists is probably derived from ferric oxides in the original shales. It is likely, therefore, that the magnetite-rich schists were originally red shales. The non-magnetitic schists, however, are more likely to have been green shales with the iron largely in the ferrous form.

The greenstone schists of the Pinney Hollow formation vary from a few inches to several hundred feet in thickness and are concordant with the interbedded sericite schists. The essential constituents are chlorite, epidote, albite and calcite or ankerite. The most conspicuous minor constituent is magnetite in small octahedra similar to those in the sericitic schists. In regions where the sericitic schists contain garnet and biotite the greenstones commonly contain a bright green, actinolitic amphibole and noticeably less chlorite and carbonate. The amphibole commonly occurs as tiny needles lying in the foliation plane and sometimes showing a lineation as well. In some occurrences, however, the amphibole occurs as rods an inch or more in length with random orientation.

The majority of the greenstones show a marked compositional banding with laminae between an eighth and a half inch in thickness. The banding is caused by variable proportions of the major constituents. Some of the bands or beds are so rich in calcite or ankerite that they might be more properly called chloritic limestones or dolomites. Most of these carbonate rich bands are only a fraction of an inch thick but beds several feet thick have been found. The banding of the greenstones and their interstratification with probable argillaceous sediments strongly suggests subaqueous deposition. A straight sedimentary origin seems unlikely unless one is willing to admit the possibility of a ferruginous, shaly dolomite. Inasmuch as the composition of the greenstones approaches rather closely that of an andesite or basalt, however, a tuffaceous origin is more probable. That the greenstones may be tuffs, or reworked tuffaceous material, is supported by the occurrence in the Saxtons River quadrangle of massive greenstones with ovoid knots of epidote, apparently amygdules, in the lower part of the Pinney Hollow formation. The writer has seen similar rocks associated with banded greenstones in the Barnard gneiss and the Shaw Mountain formation. In southern Quebec Cooke (1937) described pillow lavas and other unquestionable volcanic rocks in strata resembling the Pinney Hollow and associated formations, and quite possibly their stratigraphic equivalents. The writer believes that the Pinney Hollow greenstones are related to mafic vulcanism and are largely of pyroclastic origin.

The Pinney Hollow formation as traced southward by Rosenfeld and Skehan proves to be essentially equivalent to the Rowe schist (Emerson, 1917) of western Massachusetts, and the greenstones in its upper part to the Chester amphibolite of Emerson. The Pinney Hollow appears to be continuous to the north with the upper part of the Bennett or Sutton schists of southern Quebec (Fairbairn, 1932). The quartz-sericite-chlorite schists of the Pinney Hollow resemble, both lithologically and in stratigraphic sequence, certain phyllites in the Taconic Range mapped as the Mettawee formation by Kaiser (1945) and as Mettawee and Nassau by Fowler (1949). The Mettawee is overlain conformably by the Schodack formation of known lower Cambrian age. Prindle and Knopf (1932) correlated the Mettawee of the southern Taconic area with the Rowe east of Hoosac Mountain.

Ottauqueechee formation

The Ottauqueechee formation (Perry, 1928) is typically a coal-black, graphitic phyllite with interbedded graphitic quartzites, and including minor horizons of sericite schist and greenstone similar to those of the Pinney Hollow. The graphite on a freshly broken surface of some of the phyllites is sufficiently abundant to soil the fingers. Biotite and garnet porphyroblasts are common in regions of relatively high grade metamorphism. A minor but distinctive type in the Ottauqueechee is a rock consisting of alternating laminae of micaceous and granulose minerals averaging slightly less than an eighth of an inch in thickness. The granulose laminae are composed largely of quartz with small amounts of albite, and the micaceous laminae of sericite, chlorite and biotite.

The Ottauqueechee formation is about 1,400 feet thick in Ludlow thinning southward to about 500 feet in the southern part of Andover. It has not been identified in western Massachusetts although it may be equivalent to the lower part of the Savoy schist of Emerson. Its northward extension, however, is continuous with the Mansonville formation (Fairbairn, 1932) of southern Quebec. The writer correlates the Ottauqueechee phyllites with the black slates of the lower Cambrian Schodack formation of the Taconic Range.

Stowe formation

Cady (personal communication) has suggested that the schists immediately overlying the Ottauqueechee formation in the Montpelier quadrangle be designated the Stowe formation. The rocks correlated with the Stowe formation in the Ludlow area are equivalent to the lower half of the Bethel schist as mapped in Plymouth and the Bridgewater townships by Perry (1928). The Stowe formation is so similar lithologically to the Pinney Hollow formation that a detailed description would be a needless repetition. The Stowe formation does, however, outcrop in a region of slightly higher grade metamorphism so that the schists are more generally garnetiferous and the greenstones more generally amphibolitic than in the Pinney Hollow.

The thickness varies from about 900 feet in Ludlow to about 600 feet in Andover. The distribution of greenstones is erratic and they do not form any well-defined horizon within the formation. The contact with the Ottauqueechee is placed at the top of the uppermost graphitic schist.

The Stowe formation is correlated with the Wallace Ledge slate of the Taconic area (Kaiser, 1945; Fowler, 1949). The Wallace Ledge slates overlie the Schodack formation conformably and are lithologically like the slates of the Mettawee formation. The sequence Mettawee-Schodack-Wallace Ledge of the Taconics is remarkably similar to the sequence Pinney Hollow-Ottauqueechee-Stowe of the Ludlow area, both lithologically and in the relative thicknesses of the three units.

Moretown formation

The rocks mapped as the Moretown formation in the Ludlow area include the upper half of the Bethel schist, and the lower half of the Mississquoi group as mapped in Plymouth and Bridgewater by Perry (1928). The reasons for dropping the names Bethel and Mississquoi are that both have been used elsewhere for other formations, and that the limits of the formations were rather vaguely defined. The name Moretown has been proposed by Cady (personal communication) because of typical exposures near the village of Moretown, west of Montpelier. In the Ludlow area the central part of the formation has been mapped separately as the Whetstone Hill member. The base of the Whetstone Hill member coincides with the base of the Mississquoi group as mapped by Perry.

The upper and lower portions of the formation somewhat resemble the peculiar banded rock of the Ottauqueechee except that the granulose laminae are generally thicker than the micaceous laminae almost paper-thin, giving the rock a "pin-striped" texture (Plate VI). The granulose laminae average about a half inch in thickness but range from an eighth of an inch to an inch or more. The micaceous laminae are mineralogically similar to those in the Ottauqueechee. Garnets, when present, are usually centered on these laminae. Some of the granulose laminae are fairly pure quartzites but most contain sodic plagioclase, small amounts of epidote and scattered grains of calcite or ankerite. In higher grades of metamorphism the carbonates give way to porphyroblasts of hornblende, either randomly oriented in the granulose layer or forming rosettes, the "fasciculites" of Hitchcock and Emerson, in the micaceous layer. The fasciculite schists of New England appear to be similar to the "garbenschiefer" of German petrologists. Ordinary greenstone schists, with or without amphibole, also occur in the Moretown formation, but are relatively less abundant than in the underlying formations.

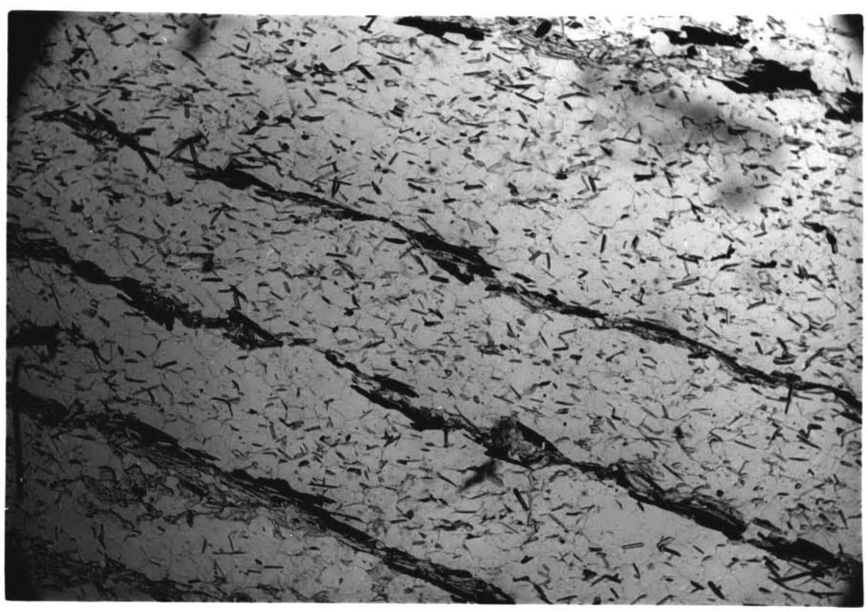
The origin of the peculiar "pinstripe" quartzites and fasciculite schists of the Moretown formation is problematic. The writer believes that the pinstripe texture is primary inasmuch as it appears to be always concordant with the bedding as shown by the large scale compositional banding. It is cross-cut in places by a secondary slip cleavage but

Plate VI

Pinstripe quartzite of Moretown formation.

The specimen is from the north side of Route 103 in Smithville. The pinstriping at this locality is parallel to the bedding as shown by large-scale compositional banding.

A



B

the two appear to be distinct. According to Hadley (1942), however, the pinstripe texture in the Albee formation of western New Hampshire is a recrystallized fracture cleavage. Cooke (1937) describes rocks from the Caldwell series of southern Quebec, apparently similar to the pinstriped quartzites of the Ludlow area, and interprets them as of tuffaceous origin. The high quartz content of some of the granulose laminae, however, is unlikely in a tuff. Harker (1932) suggests that the "garbenschiefer" and similar types may be metamorphosed calcareous grits or greywackes, but warns of their compositional similarity to common igneous rock types.

The Whetstone Hill member of the Moretown formation includes rock types similar to those occurring elsewhere in the formation but is essentially a black to grey-green, quartz-sericite-garnet-biotite schist. Plagioclase is present but as finely disseminated grains rather than the conspicuous augen characteristic of the Grahamville formation. The graphitic coloring matter is very erratically distributed in the formation and does not appear to have any stratigraphic significance. Magnetite and tourmaline are the most common accessories but magnetite has been observed only in the non-graphitic schists. A prominent feature of the schists of the Whetstone Hill member is the occurrence of flattened lenses consisting of tiny pink garnets, generally stained black by manganese oxides. The lenses average a half inch to an inch in thickness, are a foot or more wide and may be several feet long. They are flat in the foliation plane and elongate down the dip as are the quartz lenses in the Pinney Hollow. It is possible that they represent deformed and metamorphosed manganiferous concretions.

A peculiar manganiferous "ironstone" occurring in the lower part of the Whetstone Hill member may be genetically related to the garnet lenses. The ironstone, varying from three to twenty-five feet in thickness, is best exposed on East Hill in Andover and on Whetstone Hill in Ludlow. The mineral constituents are magnetite, spessartite garnet, manganese-rich carbonate, and a manganiferous, weakly magnetic, magnetite-like mineral which is probably franklinite. Quartz is variable in amount, absent in some occurrences, but in others constituting more than half the rock. One of the more siliceous varieties was formerly quarried on a small scale for the manufacture of whetstones.

The Moretown formation is lithologically similar to the Savoy schist of Massachusetts and appears to be continuous with it. The formation varies between 3,500 feet and 3,700 feet in thickness. In the Ludlow area the lower pinstripe member is about 500 feet thick, the Whetstone Hill member about 1,600 feet thick and the upper pinstripe member also about 1,600 feet thick. There is no obvious correlation between the Moretown formation and formations of the Taconic or Champlain Valley sequence. The Zion Hill quartzite (Kaiser, 1945) of the Taconics, however, is a ferruginous quartzite overlying the Wallace Ledge slates and underlying early Ordovician black slates. It is probably the most likely correlative of the Moretown formation.

Cram Hill formation and Barnard gneiss

The Cram Hill formation (Currier and Jahns, 1941) corresponds to the upper half of the Mississquoi group of Richardson. The lower boundary of the formation has not been mapped continuously between the type locality, southwest of Barre, and the Ludlow area, but it is reasonably certain that approximately the same horizon has been chosen in both areas. It is possible, however, that all, or part, of the upper pinstripe member of the Moretown is actually the equivalent of the Harlow Bridge quartzite forming the basal member of the Cram Hill in the type section. The Cram Hill formation contains no lithologic types markedly different from those already described as occurring in the preceding formations. The Cram Hill formation resembles most closely the Ottauqueechee, consisting largely of graphitic schists or phyllites interbedded with banded, graphitic quartzites. Banded greenstones, and feldspathic quartz-sericite-biotite schists, commonly showing a lamination similar to that in the pin-striped quartzites of the Moretown formation, are fairly common. The chief difference between the Cram Hill and Ottauqueechee formations is a tendency for the rocks of the Cram Hill formation to be slightly more feldspathic.

Rusty-weathering, sometimes graphitic, plagioclase-rich gneisses and feldspathic quartzites are particularly characteristic of the upper part of the Cram Hill formation in the southwest corner of Cavendish.

The thickness of the Cram Hill formation and Barnard gneiss is difficult to estimate owing to the relatively intense deformation in the vicinity of the axis of the Proctorsville syncline and the unconformable relations of the overlying Shaw Mountain formation. The uncorrected thickness at the latitude of Ludlow is about 6,000 feet and in southern Andover about 3,500 feet. The upper part of the formation, however, is probably missing in Andover.

The Barnard gneiss (Richardson, 1927) includes, as mapped by the writer, a variety of gneisses apparently stratigraphically equivalent to the schists and quartzites of the Cram Hill formation. In the region known as "The Alps" in the southwest part of Reading and the northwest part of Cavendish, the Barnard gneiss is the only facies present. West of Proctorsville, however, schists appear in the lower part of the formation, and farther south the gneisses disappear entirely.

The gneisses include biotite gneiss, biotite-hornblende gneiss, hornblende gneiss, garnetiferous hornblende gneiss and epidote amphibolite. The essential constituents are quartz, plagioclase (albite-oligoclase), biotite, hornblende and sometimes almandite garnet. Minor constituents include clinozoisite, magnetite, chlorite, muscovite and calcite. In some occurrences calcite is sufficiently abundant to rank as a major constituent.

The plagioclase generally shows combined albite and pericline twinning and contains numerous inclusions of sericite and epidote, similar to those in the plagioclase feldspars of the Green Mountain basement rocks. The inclusions often show a zonal arrangement and are most abundant in the centers of the grains, the borders of the grains being relatively clear. In some of the most massive gneisses the plagioclase grains are almost euhedral and occur in a relatively fine-grained matrix. The hornblende is a strongly pleochroic variety with X yellow, Y green and Z a deep blue green.

Inasmuch as most of the lithologic types in the Barnard gneiss lie within the compositional range of common igneous rocks the writer has interpreted them as of probable volcanic origin. The well-bedded types could thus be metamorphosed pyroclastics, and the more massive gneisses and amphibolites possibly flows or sills, ranging from dacite to basalt in composition. That some of the amphibolites may indeed be flows is supported by the occurrence in them of small ellipsoidal masses of epidote strongly resembling amygdules. Evidence indicating that some of the massive biotite gneisses may be flows or sills is the occurrence of subhedral feldspar grains in a relatively fine-grained groundmass. It seems unlikely that plagioclase grains which have been altered to the extent of those in the Barnard gneiss would have been durable enough to have survived any ordinary sedimentary process in such abundance, and show, at the same time, subhedral outlines. Although similarly altered feldspars are common in the Green Mountain basement rocks, few appear as detrital grains in the overlying arkoses and greywackes. Another feature making a straight sedimentary origin unlikely is the suddenness of the facies change between the Barnard gneiss and the Cram Hill schists and quartzites.

The Cram Hill formation and Barnard gneiss are apparently equivalent to the Hawley formation of Massachusetts, which consists largely of amphibolite, hornblendic gneiss and fasciculitic hornblende schist. . The most probable correlation with the Taconic sequence appears to be with the Normanskill formation (Black River to lower Trenton). The Normanskill formation (Ruedemann, 1930) consists of black shale, radiolarian chert and grit. The grits are apparently facies-equivalents of the cherts and shales, suggesting the relations between the Cram Hill schists and quartzites and Barnard gneiss of the Ludlow area. It is tempting to consider the Normanskill grits as of possible volcanic origin, and significant in this respect that the bentonites, characteristic of rocks of Black River age elsewhere in the Appalachian region, are interpreted as altered ash beds.

<u>Shaw Mountain formation</u>

The Shaw Mountain formation, according to Currier and Jahns (1941), overlies the Cram Hill formation unconformably. In the present area the Shaw Mountain formation has been observed in contact with only the Barnard gneiss and the relationships, in each instance, are apparently conformable. The conglomeratic nature of the base of the Shaw Mountain, however, and the striking lithologic differences between the rocks above the Shaw Mountain horizon and those below, are probably sufficient indications that the horizon marks an important stratigraphic break. Currier and Jahns have traced the base of the Shaw Mountain formation northward to the southern tip of Lake Memphremagog, where it rests, according to Cady (personal communication), upon schists of the Stowe formation. To the south the Shaw Mountain unconformity is continuous with the base of the Goshen schist as mapped by Emerson (1917). At the Massachusetts-Vermont state line the Goshen rests upon schists of the Hawley (Cram Hill-Barnard gneiss) formation and farther south upon schists of the Savoy (Moretown) formation.

On the western limb of the Proctorsville syncline the Shaw Mountain formation consists of quartz-conglomerate and quartzite at the base overlain by silvery, highly garnetiferous, muscovite schists interbedded with banded epidote-amphibolites, locally rich in carbonate and chlorite. The pebbles in the conglomerates are mostly of quartzite and are shaped like slightly flattened cigars, elongate parallel to the fold axes and flattened in the plane of schistosity. The formation has a maximum total thickness of about five hundred feet, but thins both northward and southward and has not been observed north of the Plymouth-Reading road or south of Proctorsville.

Crinoid stems have been found by Currier and Jahns (1941) in limestones of the Shaw Mountain formation near Barre, the size and development of the stems indicating that the limestone is at least as young as Middle Ordovician. The overlying Waits River limestones, however, have recently yielded definite Upper or Middle Ordovician fossils (Cady, personal communication), thus limiting the age of the Shaw Mountain. It seems reasonable, therefore, to equate the base of the Shaw Mountain formation with similar breaks occurring in the Ordovician of the Taconic and Champlain Valley sequences. In the Champlain Valley, this is the unconformity at the base of the Hortonville slate, and in the Taconic Range, that at the base of the Rysedorph conglomerate.

Northfield and Waits River formations

The interbedded limestones and graphitic phyllites of the Northfield and Waits River formations overlie the Shaw Mountain formation with apparent conformity. The same distinctive lithologic types are present in both the Northfield and Waits River, but in different proportions. The dominantly argillaceous Northfield is the "clay slate" of Hitchcock (1861) and the dominantly calcareous Waits River is his "calciferous mica schist".

The typical phyllites are dark grey to black quartz-sericite schists with, in most occurrences in the Ludlow area, abundant dark red garnets occurring as perfect dodecahedra 4 - 6 mm. in diameter. The argillaceous rocks of the Northfield and Waits River are more highly garnetiferous and less feldspathic than the dark schists of the Cram Hill in the same grade of metamorphism, and the graphitic coloring matter more evenly distributed in a finer, more phyllitic matrix. Although sometimes rusty-weathering, outcrops of the Northfield phyllites do not show, in the Ludlow area, the yellow-green sulphate stain characteristic of the graphitic schists of the Ottauqueechee and Cram Hill. The garnetiferous schists of the Shaw Mountain formation are the most similar, texturally, but contain little or no graphite. The proportion of mica to quartz is distinctly higher than in any of the pre-Shaw Mountain argillaceous rocks, and the quartz lenses typical of most of the older formations are rare except in some of the basal beds of the Northfield.

The limestones are grey or blue-grey on a fresh surface but weather to a rusty, porous crust which may be several inches thick. The major constituents are calcite and quartz in variable proportions, with scattered flakes of sericite and, in some occurrences, small rhombs of rusty-weathering ankerite about a millimeter in diameter. Zoisite and lime-garnet are fairly common constituents of some of the more schistose limestones. The garnets are poekiloblastic with inclusions of calcite and quartz, and are poorly formed, having the appearance of large warts on weathered surfaces of the limestone.

The Northfield and Waits River formations have not been differentiated on the map although there is a definite zone 500 - 700 feet thick, immediately overlying the Shaw Mountain formation, which is dominantly phyllitic. This apparently corresponds to the Northfield formation as mapped by Currier and Jahns farther north, although much thinner. Inasmuch as both formations contain the same lithologic types in different proportions, however, the line of demarcation is fuzzy at best.

In Massachusetts the Northfield and Waits Riversare mapped as Goshen and Conway, respectively. The thickness of the Northfield and Waits River in the Ludlow area is at least 5,500 feet, but probably only the lower portions of the Waits River formation are present. Outside the area mapped by the writer the Waits River formation is overlain by the Gile Mountain formation (Leyden argillite in Massachusetts). The Gile Mountain formation like the Northfield, is relatively non-calcareous.

Richardson (1924) describes graptolites from slates of the Northfield formation which were identified by Ruedemann as middle Ordovician, possibly Normanskill. The validity, as fossils, of Richardson's graptolites has been questioned by Foyles (1930) and by Currier and Jahns (1941), but the occurrence is still listed as valid by Reudemann in his "Graptolites of North America" (1947, p. 63). Cady, however (personal communication) has more recently found corals of definite Middle or Upper Ordovician age in the Waits River formation, indicating that, whether valid or not, the age determination based upon Richardson's graptolites was not too far from the truth. C. G. Doll (1943) has presented evidence that the upper part of the Waits River formation and the conformably overlying Gile Mountain formation may be as young as Silurian or Devonian, but the validity of Doll's fossils has been questioned by W. S. White (1946). That as homogeneous a lithologic unit as the Waits River formation could be Middle Ordovician near its base and as young as Silurian or Devonian in its upper part, however, seems highly unlikely, particularly in view of the major unconformity separating Ordovician from Silurian or younger strata elsewhere in northeastern United States and southeastern Canada.

As has been stated previously, it seems reasonable to equate the unconformity at the base of the Shaw Mountain with the similar unconformities occurring in the Middle Ordovician of the Taconic and Champlain Valley sequences. This would make the Shaw Mountain formation the approximate equivalent of the Rysedorph conglomerate and Tackawasick limestone of the Taconic sequence (Reudemann,

1930) and possibly the Whipple marble of the Champlain Valley sequence (Fowler, 1949), of mid-Trenton age. The Northfield and Waits River formations would then be equivalent to the Snake Hill shales of the Taconics and the Hortonville (Canajoharie) shale of the Champlain Valley, a correlation that seems reasonable on lithologic as well as paleontologic and stratigraphic grounds, although the Waits River is somewhat more calcareous than might be expected from extrapolation of the normal east-west facies changes in the Trenton strata eastward across the Green Mountains.

Validity and significance of the correlation

The tentative correlation of the strata of the Ludlow area with those of other parts of the Green Mountain-Taconic-Champlain Valley region has been summarized in Figure 4. The non-fossiliferous nature of the pre-Shaw Mountain strata decreases, of course, the certainty of their correlation, but the writer feels that certain of the lithologic correlations, in particular the correlation of the Tyson and the Grahamville formations with the Mendon series, Cheshire quartzite and Rutland dolomite of the Champlain Valley sequence; and the correlation of the Pinney Hollow, Ottauqueechee and Stowe with the Mettawee, Schodack and Wallace Ledge of the Taconics, are not only plausible but probable. There is, in fact, considerable evidence (P. H. Osberg and W. M. Cady, personal communication) that the conglomerates and greywackes of the Tyson formation

Probable Age	Champlain–Vermont Valley: So. Vermont - W. Mass.	Champlain–Vermont Valley: West-central Vermont	Taconic Range	Eastern Green Mountains: Eastern Berkshires	Eastern Green Mountains: Eastern Vermont	Eastern Green Mountains: Southeastern Quebec
Late Middle Ordovician	Walloomsac sl.	Hortonville sl. Whipple ls.	Snake Hill sh. ? Tackawasick ls Rysedorph cgl.	Leyden argillite Conway sch Goshen sch.	Gile Mtn Fm Waits River Fm. Northfield sl. Shaw Mtn Fm	St Francis series (= Tomifobia) Beauceville fm (= Magog sl)
— Unconformity —						
Early Middle Ordovician	Stockbridge ls.	Glens Falls ls. Orwell ls. Middlebury ls. Beldens fm. Crown Point ls.	Normanskill fm.	Hawley sch	Barnard gneiss Cram Hill fm. Missisquoi gp.	Caldwell series
Lower Ordovician	Stockbridge ls.	Bridport dol. Bascom fm. Cutting dol. Shelburne marble	Bald Mtn ls. Deepkill sh. Schagticoke sh.	Hawley sch	Cram Hill fm. Missisquoi gp.	Caldwell series
Middle and Upper Cambrian	Stockbridge ls.	Clarendon Springs dol. Danby fm.	Zion Hill qte	Savoy sch.	Moretown fm. Missisquoi gp.	Caldwell series
Lower Cambrian	Stockbridge ls. Cheshire qte. Dalton fm. Vermont fm.	Winooski dol. Monkton qte. Dunham dol. Rutland dol. Cheshire (Gilman) qte. W. Sutton sl. (= Moosalamoo) White Bk. dol. (= Forestdale mbl) Pinnacle ark. (= Mickwaukett) Mendon ser.	Wallace Ledge sl. Schodack sl. + ls Mettawee sl. Bomoseen grit Nassau fm.	Savoy sch. Chester amphibolite Rowe sch. Hoosac sch. Dalton fm.	Stowe fm. Bethel sch Ottauqueechee fm. Pinney Hollow fm. Grahamville fm. Tyson fm.	Mansonville fm. Bennett (= Sutton) sch

Figure 4. Correlation Chart

are continuous around the northern end of the Green Mountain anticlinorium with those of the Mendon series.

It is probable that the Tyson and Grahamville formations might also be correlated to advantage with the pre-Mettawee of the Taconic sequence, but the stratigraphy of the pre-Mettawee of the Taconics is somewhat confused in the literature and a detailed correlation was not attempted. The Mettawee and younger Taconic formations differ from the Pinney Hollow and younger formations of the Ludlow area in their generally much lower grade of metamorphism and in the presence of a few fossiliferous horizons. More critical differences, however, are the relative absence of greenstones, and in the occurrence of relatively thin horizons of limestone and dolomite, important in that they contain most of the fossils, in that part of the sequence between the base of the Mettawee and the horizon of the Rysedorph conglomerate. Inasmuch as the probable equivalent strata of the Ludlow area are rich in greenstones and almost devoid of carbonates, other than some of the impure dolomitic rocks associated with some of the greenstones, the Taconic sequence appears to be intermediate in this respect between the Champlain Valley sequence and that of the Ludlow area. If the eastward facies change in the Cambro-Ordovician of western New England is at all regular, as seems probable, this would mean that the original site of deposition of the now allocthonous Taconic sequence would have been somewhere between the other two. This in turn, would suggest that the allocthone must "root" somewhere in the region of the Green Mountain anticlinorium.

Assuming that the correlations are essentially correct, the east-west facies change in the pre-Shaw Mountain formations would be from the argillaceous sediments and volcanics of the Ludlow area, to the argillaceous sediments with minor volcanics and carbonates of the Taconic sequence, to the carbonates of the Champlain Valley sequence. It is tempting to speculate that the argillaceous sediments and carbonates may be ultimately derived from volcanic material. It is significant, in particular, that the dolomitic carbonate rocks occur chiefly in that part of the Champlain Valley sequence presumably equivalent to the Pinney Hollow, Ottauquechee, Stowe and Moretown formations, in which greenstones are particularly abundant. A rock made of ferruginous shale and dolomite in the proper proportions would have essentially the bulk composition of an andesite or basalt plus carbon dioxide and water. The potash-soda ratio would be a bit too high, but if some sodium were assumed to have entered the sea water, the discrepancy would be less critical, and it is furthermore a characteristic of early Paleozoic shales of northeastern United States that they have a relatively high ratio of soda to potash.

According to Cady (1945) the basal clastic facies of the Champlain Valley sequence is derived from a western source, presumably the Adirondacks. The pebbles of the Tyson conglomerates, however, appear to be of relatively local origin, derived largely from the quartzites of the Ludlow Mountain-Sawyer Rocks belt.

Stratified Rocks of the Chester Dome

Reading gneiss

The gneissic rocks in the core of the Chester dome were named the Reading gneiss by C. H. Richardson (1928). Although there is considerable lithologic variation within the gneisses, it has not been possible to map the different types separately. The most common types are, in approximate order of decreasing abundance, biotite gneiss, biotite-muscovite gneiss, biotite amphibolite, and feldspathic muscovite-biotite schist. The compositional bands vary from several feet to a fraction of an inch in thickness. The more thinly-banded gneisses are most characteristic of the marginal regions of the dome and of the gneisses adjacent to the marble and schist of the Hawks Mountain area.

Biotite gneiss is the dominant rock type, particularly among the more coarsely banded and coarser-grained gneisses. The major mineral constituents, in usual order of abundance, are: quartz, plagioclase (oligoclase), biotite, clinozoisite, and lesser amounts of microcline, muscovite or hornblende. The accessory constituents include magnetite, tiny red garnets and sphene. Sphene and hornblende, however, have not been found in any of the muscovitic gneisses whereas garnet tends to be relatively more abundant in these rocks. Microcline, although minor or absent in most of the gneisses, is, in some, the dominant feldspar. Where abundant, microcline commonly occurs as pink augen averaging approximately

1 x 3 cm. in cross-section.

The biotite amphibolites commonly occur as minor concordant bands in the other gneisses, varying from a few inches to several feet in thickness. The mineral constituents include: strongly-pleochroic amphibole, biotite, quartz, plagioclase (oligoclase-andesine) and epidote, with minor amounts of sphene and magnetite.

The biotite-muscovite gneisses are quite similar to the biotite gneisses except for their higher muscovite content. Highly muscovitic gneisses, however, are most typical of the more thinly-banded portions of the Reading gneiss. Another common type in the thinly-banded gneisses is a schistose gneiss or feldspathic schist containing quartz, muscovite, plagioclase and biotite with minor amounts of clinozoisite and garnet. In some occurrences these gneisses contain as much as five or ten per cent garnet. Less common types include quartz-muscovite-garnet schist, quartzite, feldspathic quartzite and garnetiferous quartzite.

That some of the Reading gneiss is of sedimentary origin seems indisputable. The relatively non-feldspathic schists and quartzites are probably metamorphosed shales and sandstones, respectively. The more feldspathic schists and schistose gneisses lie in the compositional range of siltstones or of argillaceous sediments contaminated by felsic pyroclastics. The more feldspathic gneisses might be interpreted as metamorphosed arkoses, or siltstones low in argillaceous material, or as felsic pyroclastics. The biotite amphibolites might

Figure 5.

Stratigraphic Column, Chester Dome

Formation	Thickness	Lithology	Age
Northfield and Waits River	--	Garnetiferous, graphitic schist, blue-grey siliceous limestone, coarse amphibolite	Middle Ordovician
Shaw Mountain	0 - 400'	Quartz-conglomerate, quartzite, cummingtonite quartzite, quartzite, garnet-muscovite schist, impure marble, epidote amphibolite.	" "
---------Unconformity---------			
Barnard gneiss		Biotite and hornblende gneiss garnetiferous hornblende gneiss, amphibolite, fasciculitic hornblende schist	" "
Cram Hill formation		Graphitic schist, quartzite, some epidote amphibolite	" "
Moretown formation	1,000'	Pinstripe quartzite, feldspathic, muscovite-garnet schist, fasciculitic hornblende schist	" "
Stowe formation		Garnetiferous quartz-muscovite-chlorite schist, banded epidote-amphibolite	Probably lower Cambrian
Ottauquechee formation		Graphitic schist, quartzite banded epidote-amphibolite	" " "

Figure 5 - Continued

Formation	Thickness	Lithology	Age
Pinney Hollow formation	1,000'	Garnetiferous quartz-muscovite-chlorite schist, banded epidote-amphibolite, staurolite-kyanite schist	Probably Lower Cambrian
Grahamville (?) formation	700'	quartz-muscovite-biotite-plagioclase schist and schistose gneiss, impure quartzite, biotite-amphibolite	" " "
Bull Hill gneiss	0-1000'	Microcline augen-gneiss	" " "
Star Hill formation	0-200'	Quartz-muscovite-biotite-albite schist, garnetiferous mica schist, kyanite-staurolite schist, calcite and dolomite marble.	" " "
		-----------Angular Unconformity------------	
Cavendish schist	?	Quartz-muscovite-biotite-albite schist, garnetiferous mica schist, kyanite-staurolite schist	Pre-Cambrian
Whitesville marble	0-300'	Dolomite marble, Ca-Mg silicate rock, minor beds of quartzite and schistose muscovite-biotite gneiss	" "
Reading gneiss	?	Biotite gneiss, biotite-muscovite gneiss, biotite amphibolite and schistose muscovitic gneiss	" "

correspondingly be interpreted as either metamorphosed calcareous greywackes or as intermediate (andesitic or basaltic) pyroclastics. On the other hand, the "igneous" compositions of the more feldspathic gneisses and of the biotite amphibolites would not be consistent with the interpretation of these rocks as banded intrusives, concordant with the interlaminated metasediments. The most likely candidates for such an origin would probably be the relatively coarse gneisses in the central part of the dome. These gneisses have compositions approaching those of typical granodiorites or quartz-diorites. They do not, however, show anywhere a clear cross-cutting relationship to the other lithologic types, and the banding within them is invariably parallel to that in the surrounding rocks. There is no evidence, other than that of a plausible bulk composition, that they are of igneous origin.

Whitesville marble

The Whitesville marble is in apparent conformable contact with the Reading gneiss and, on the assumption that the Hawks Mountain complex is synclinal, is assumed to overlie the gneiss. The marble, chiefly dolomitic, also includes a varied assortment of calcium-magnesium silicate rocks as well as bands of gneiss and quartzite identical with the adjacent, banded phase of the Reading gneiss. These may be in part infolds but most are believed to represent a primary interstratification of gneiss and marble. One particularly persistent band, averaging about fifty feet thick, seems to occur near the center of the formation along most of the northwestern margin of the Hawks Mountain complex. In areas where it is reasonably certain that there is not much duplication by folding, the formation varies from zero to about 300 feet in thickness. The variability may be primary but is more likely the result of plastic flow in the carbonate beds.

The dolomitic marbles, formerly quarried extensively for lime in Amsden and Cavendish, are commonly white to buff-colored when pure. Most occurrences, however, contain silicates in varying amounts. It is interesting that the calcite-dolmite ratio varies almost directly with the amount of silicate, suggesting that the calcite is probably a by-product of the formation of the dominantly magnesian silicates. The chief silicates are phlogopite, tremolite or actinolite, diopside, and less commonly epidote. Some of the more distinctive rock types include: tremolite-phlogopite schist and massive tremolite-diopside rock. Epidote occurs chiefly along the contacts between the marbles and the more schistose portions of the Reading gneiss and also along the contacts with the Cavendish schist. At the contact of the Cavendish schist the epidote is commonly associated with a dark green, actinolitic amphibole and with biotite, the whole forming what appears to be an irregular reaction zone between the schist and marble, varying from half an inch to three or four inches in thickness. Some of the marbles contain sparsely disseminated grains of sulphide, largely pyrrhotite but with some chalcopyrite.

Cavendish and Gassetts schists

The Cavendish schist (Richardson, 1928) is believed to overlie conformably the Whitesville marble. The characteristic type adjacent to the Whitesville marble is a quartz-muscovite-plagioclase (generally oligoclase)-biotite schist. Toward the center of the supposed recumbent syncline this type gives way to a highly garnetiferous and less biotitic schist, this in turn to a non-feldspathic quartz-muscovite-garnet schist, and finally to a beautiful quartz-muscovite-garnet-staurolite-kyanite schist. The latter type is well known from the excellent road-cut and quarry exposures near Gassetts from which it was named the Gassetts schist by Richardson (1928). The Gassetts and Cavendish schists have not been differentiated, however, on the geologic map.

Both the Cavendish and Gassetts schists are unusually coarse-grained, the plagioclase and garnet porphyblasts commonly attaining a diameter of three-quarters of an inch and the staurolite and kyanite blades as much as three inches long in some occurrences. Several occurrences of staurolite and kyanite in parallel growth were observed. The micas occur in coarse flakes up to a quarter inch in diameter. Partial analyses of the muscovite from the Gassets schist (Currier, 1934) show it to be a highly sodic variety approaching paragonite in composition, with a sodium potassium ratio of about 1:1. The garnet is close to almandite in composition and in some instances is rimmed with chlorite. The chlorite, a relatively magnesian variety, also occurs in isolated porphyroblasts apparently unrelated to garnet. The schist at Gassetts was formerly quarried for its mica and garnet content but the mill and quarry are now closed. Minor constituents of this schist include tourmaline, magnetite, rutile, zircon, and epidote. Epidote, however, is generally limited to the plagioclase-bearing schists.

Star Hill formation

The Star Hill formation is a highly distinctive unit, although including a variety of lithologic types. In the Star Hill area the formation is essentially a garnet-muscovite schist about 200 feet thick, underlain by 5 - 10 feet of white dolomite, and including a 20 foot bed of pink calcite marble at approximately the center of the formation. Some of the schist is albitic but most is feldspar free. Kyanite and staurolite are locally abundant in the Star Hill area. South of Star Hill and in the western part of Springfield, the rocks mapped as the Star Hill formation are muscovite-biotite-albite-garnet schists associated with quartzite, feldspathic quartzite and thin beds of calcite and dolomite marble. The development of silicates in the marbles is similar to that in the Whitesville formation except that some of the calcite marbles in the Star Hill formation are relatively free of silicates, and appear to be primary rather than the result of de-dolomitization.

Bull Hill gneiss

The Bull Hill gneiss (Richardson, 1931) is a biotite-muscovite gneiss characterized by large augen of microcline (Plate VII). It varies in thickness from zero to about one thousand feet and, where both are present, appears to overlie the Star Hill formation conformably. The Bull Hill gneiss, in turn, is overlain with apparent conformity by schists and gneisses correlated with the Grahamville formation.

Although Richardson (1931) considered the Bull Hill gneiss an igneous rock, the writer considers this interpretation unsatisfactory because of the parallelism, over long distances, between the Bull Hill gneiss and horizons of schist, gneiss and marble, of definitely sedimentary origin, and because of the lack of intrusive contacts. In some areas the matrix of the augen gneiss is decidedly schistose, the rock resembling, except for the relatively large size of the feldspars, some of the pebbly greywackes of the Tyson formation. Although no trace of original grain boundaries exists in the feldspars, the writer has tentatively interpreted the Bull Hill gneiss as a metamorphosed arkose or greywacke, the large feldspar augen possibly having been derived from recrystallization of original detrital feldspars. The Bull Hill gneiss

Plate VII

Bull Hill gneiss

The locality is on the north bank of the Middle Branch of the Williams River, about one mile west of the center of Chester village. The approximate scale may be judged from the vegetation in the foreground. The augen are of pink microcline, the largest measuring about 3/4 x 1 1/2" in cross-section. The joints perpendicular to the plane of the photograph are about normal to the local fold axes.

Plate VII

and the Star Hill formation are correlated, with reservations, with the Tyson formation of the Green Mountain areas, although the Bull Hill gneiss has been included with the overlying Grahamville (?) formation on the map (Plate II).

Grahamville (?) formation

The rocks mapped as Grahamville formation in the Chester dome include muscovite-biotite-plagioclase schists and banded, schistose gneisses, resembling very much the banded phase of the Reading gneiss, but correlated, on stratigraphic grounds, with the Grahamville formation of the Green Mountain area. The formation as mapped, averages about 700 feet in thickness and appears to be conformable with both the underlying Bull Hill gneiss and the overlying Pinney Hollow schists.

Pinney Hollow and younger formations

The base of the Pinney Hollow formation is the lowest horizon in the Chester dome which can be correlated with certainty with any horizon on the eastern limb of the Green Mountain anticlinorium. The Moretown formation is the lowest unit that can be traced continuously across the axis of the Proctorsville syncline (Rosenfeld, personal communication), but the correlation of the Pinney Hollow, Ottauqueechee and Stowe formations, by their position immediately beneath the Moretown and by their lithologic similarity to the same formations in the Green Mountain area, seems straightforward.

The Pinney Hollow formation in the Chester dome includes quartz-sericite-chlorite-garnet schists, locally containing staurolite and kyanite, and, in some occurrences, albitic. The schists are interbedded, in the upper part of the formation, with banded epidote-amphibolites. The schists differ from the Pinney Hollow schists of the Green Mountain area in containing less chlorite and more garnet, and in the presence of staurolite and kyanite. The chlorite in the Chester dome schists is a more magnesian variety. Magnetite is still the chief accessory, but somewhat less abundant. The epidote amphibolites differ from the amphibolitic greenstones of the Green Mountain area in containing almost no chlorite and much less carbonate. The carbonate remaining is largely calcite rather than ankerite. In the Ludlow quadrangle there is a distinct break between the grade of metamorphism in the Pinney Hollow formation of the Chester dome and that of the Green Mountain area. In the southern part of the Saxtons River quadrangle, however, the metamorphic grade in the Green Mountain area equals that in the part of the Chester dome immediately opposite, adding certainty to the correlation.

The Ottauqueechee formation is represented by graphitic schists with interbedded quartzites. The quartzites, however, are relatively less abundant than in the Green Mountain area. The Stowe formation is almost identicalto the

Pinney Hollow lithologically but so thin that it could not be positively identified on all traverses.

The Moretown formation includes impure quartzites, with the pin-striping less pronounced than in the lower grade rocks, and quartz-sericite-biotite-garnet-chlorite-plagioclase schists which are sometimes graphitic. Schists and quartzites with fasciculitic amphiboles are somewhat more common than in the Green Mountain area, and the garnet lenses similar but coarser-grained.. The Cram Hill formation and Barnard gneiss are both present and show the same interfingering relationships as in the Green Mountain area. The Cram Hill formation is largely graphitic schist and quartzite and the Barnard gneiss includes hornblende gneiss, hornblende-garnet gneiss, biotite gneiss and fasciculitic schists. The Barnard gneiss occurs as a thin band overlying the typical Cram Hill schist around the northern end of the dome, but replaces the schists entirely as the formation passes through the western part of Springfield township. Farther south, however, in the Saxtons River quadrangle, the graphitic schists and quartzites again become the dominant facies. It is interesting that north of Proctorsville, on the western limb of the Proctorsville syncline, the formation is largely the Barnard gneiss facies, and on the eastern limb largely the graphitic schist-quartzite facies, suggesting that the Proctorsville syncline must extend to considerable depth to provide space for the facies change to take place.

Although the Shaw Mountain formation was identified at several points about the northern end of the Chester dome, occurrences were too limited in extent to be mapped. In the southwest part of Springfield, however, the formation attains a thickness of approximately four hundred feet. The lithologic types include quartz-conglomerate, quartzite, quartz-muscovite-garnet schist, and amphibolite. Types which are not found elsewhere in the formation include cummingtonite quartzite associated with the basal quartz-conglomerates, and impure calcite marbles at about the center of the formation. The cummingtonite quartzites commonly contain spessartite garnet and have a dark stain from the weathering of manganese-rich carbonates. In some instances, however, the black color of these quartzites is caused by graphite. Northeast of Litchfield Hill in Springfield the cummingtonite quartzites include a massive amphibolite averaging about ten feet thick, and containing ellipsoidal masses of epidote about a half inch long. The epidote nodules have been interpreted as amygdules and the amphibolite as a possible flow.

The phyllitic zone at the base of the Northfield-Waits River sequence around the dome is identical with that on the western limb of the Proctorsville syncline but generally less than 200 feet in thickness. A coarse amphibolite in the lower part of the Waits River formation in the extreme northeast corner of the Ludlow quadrangle may possibly be derived from the metamorphism of a dolomitic zone in the Waits

River limestones inasmuch as many of the limestones nearby contain a higher percentage than usual of the small, rusty-weathering ankerite rhombs.

The most remarkable feature of the younger rocks in the Chester dome is their extreme thinness as compared with the Green Mountain area. In the Chester dome the Pinney Hollow-Cram Hill sequence has a thickness varying between 1,000 and 1,500 feet as compared with 9,700 feet to 15,200 feet for the same sequence in the Green Mountain area, a factor of 1:10. As nearly as can be determined the various formational units are continuous, however, and have about the same relative thickness as in the Green Mountain area. Although the possibility that the thinness of the formations in the Chester dome area may be partly primary cannot be ruled out, the writer believes it to be largely the result of tectonic thinning. This hypothesis is supported by the extensive development of boudinage in the more competent beds and by the flattening, normal to the bedding, of the pebbles in the Shaw Mountain conglomerate. Because of the thinness of the formations, however, the Pinney Hollow-Cram Hill sequence has been mapped as one unit in the Chester dome area.

Correlation of the pre-Pinney Hollow rocks of the Chester dome

The correlation of the pre-Pinney Hollow stratified rocks of the Chester dome with those of the Green Mountain area is still uncertain. That the plagioclase schists, impure quartzites and schistose gneisses immediately beneath the

Pinney Hollow are to be correlated with the Grahamville formation seems probable, but where to draw the line separating the Grahamville formation from the Tyson formation or that at the base of the Tyson formation, is far from evident. The writer believes that the Whitesville marble, Cavendish schist, and the bulk of the Reading gneiss belong to the basement complex. This hypothesis is supported by the discordance between the Hawks Mountain complex and the over-all structure of the dome, and by the non-appearance of the Hawks Mountain sequence about the periphery of the dome. The Hawks Mountain sequence is tentatively correlated with the schist-dolomite-gneiss sequence in the vicinity of the Devils Den in the central part of the Wallingford quadrangle.

The banded, relatively schistose portion of the Reading gneiss, however, is so similar lithologically to the presumed equivalent of the Grahamville and Tyson formations that the writer has been unable to locate satisfactorily any line of unconformity between them, although fairly certain that such a line must exist. The safest criterion for correlation of any given horizon with the younger rocks is probably that of parallelism over long distances with the base of the Pinney Hollow. By this criterion the Star Hill formation and the Bull Hill gneiss have been correlated, tentatively, with the Tyson formation. Both the Bull Hill gneiss and the Star Hill formation contain lithologic types likely to occur in the Tyson formation in the metamorphic grade of the Chester dome, but not in the expected order. In the Tyson formation of the Green Mountain area the schists and marbles overlie, rather than underlie, the arkose-greywacke zone. Because of these difficulties the correlation of the pre-Pinney Hollow strata and the location assigned to the unconformity in the dome must be regarded as tentative. It is quite possible, however, that more intensive field work in the Chester dome area might solve the problem. Unfortunately, time was not available for further study of the key areas before the preparation of this report.

Unstratified rocks

Ultramafics

Altered ultramafics are of common occurrence in the Ludlow area. The bodies range in size from small sill-like or pod-shaped masses a few feet across, to stock-like bodies a half mile to a mile in diameter such as that west of Proctorsville.

The larger bodies, specifically those in the Proctorsville area and the one west of Chester village, are chiefly of serpentine bordered or veined by talc-carbonate rock. All of the smaller bodies consist entirely of talc-carbonate rock. Although some of the serpentines elsewhere in Vermont contain zones of relatively unaltered dunite or peridotite (Bain, 1936), no unaltered olivine or pyroxene has been observed by the writer in any of the serpentines of the Ludlow area. In the central part of the Proctorsville serpentine, however, some of the antigorite occurs in block-like masses 2 or 3 mm. across. It is possible that these are pseudomorphs after olivine.

The serpentines consist almost entirely of antigorite with magnetite and chromite as the chief accessories. A few small veinlets of cross-fiber chrysotile were observed in the Proctorsville body but none of the fibers were more than 2 mm.long. The talc-carbonate rocks consist largely of talc and dolomite or ankerite, although magnesite also occurs at Chester. At the old soapstone quarry west of Perkinsville, and at Chester, much of the talc in the soapstone has a fibrous habit, commonly occurring as small rosettes about a half inch in radius. The writer interprets the fibrous talc as pesudomorphous after anthophyllite. The talc-carbonate rocks are apparently an alteration of the serpentine inasmuch as the characteristic textures of the serpentine commonly continue, without interruption, across the contact. Many of the smaller talc-carbonate bodies, now containing no serpentine, show these textures, probably as a relict feature.

At the immediate contacts of the ultramafics with the wall rocks, odd rock types have been observed. In the Green Mountain area these are commonly chlorite, chlorite-ankerite or talc schists. At several points about the border of the Proctorsville serpentine, these types have been observed in association with actinolite rock and actinolite-chlorite rock. In the ultramafics of the Chester dome, the preceding types occur and also biotite schists, biotite-chlorite schists, chlorite schists, talc-actinolite rock and massive talc. These rocks tend to be arranged in concentric zones. At Chester a typical sequence, starting at the wall rock, is: biotite schist, chlorite schist (with magnetite and pyrite), talc-actinolite rock, talc, talc-carbonate rock, serpentine. The talc quarry at Chester is well known to mineral collectors for the fine specimens of actinolite, chlorite, talc, magnetite, and pyrite that may be obtained there.

The larger ultramafics, in particular the Proctorsville body, are quite clearly cross-cutting in their relations to the wall rock. Many of the smaller bodies, however, are rather remarkably concordant, and, even though not more than a few tens of feet thick, may maintain continuity along strike for a mile or more. A series of these tabular soapstone bodies occurs along the Moretown-Cram Hill contact south of Smithville and may all possibly be a part of one large sheet. Each of these bodies appears to occur at precisely the same horizon. The underlying rock in each instance is a greenstone, resting upon pin-striped quartzites of the Moretown formation, and the overlying rock a black schist of the Cram Hill formation. It is interesting that even in tabular bodies of this sort occurring at other horizons the foot wall is commonly greenstone, suggesting some genetic relationship between the greenstone and the talc-carbonate rock. The association of greenstone with ultramafics has also been noted by Hawkes (1941, p. 92.)

Although the Vermont ultramafic belt has received considerable attention (Gilson, 1927; Bain, 1936-1942; Hess, 1933; Phillips and Hess, 1937), the date of their emplacement is still uncertain. In the Ludlow area they appear to antedate most of the deformation and metamorphism in that they have numerous minor structural features in common with the enclosing schists, and in that their alteration products appear to be related to the local grade of metamorphism. It is noteworthy that nowhere in Vermont or western Massachusetts have ultramafics yet been found in rocks above the Shaw Mountain unconformity. It has been suggested that this is because the Shaw Mountain and younger formations outcrop too far to the east of the ultramafic belt, but where older rocks appear east of their normal belt of outcrop, however, as in the Chester dome, they still contain ultramafics. This would suggest that the so-called "serpentine belt" may have more stratigraphic control than is generally realized. Positive evidence on the age of the ultramafics has been provided by Clark and Fairbairn (1936) who report serpentine pebbles in a basal Silurian breccia near Lake Memphremagog, and by Bain (1942) who reports serpentine pebbles in an Ordovician conglomerate near Trowsers Lake, Quebec. The conglomerate at Trowsers Lake is apparently that at the base of the Beauceville formation and therefore probably equivalent to the Shaw Mountain. This agrees well with the Vermont-Massachusetts evidence but not with Cooke (1937) who states that serpentines occur in the basal part of the Beauceville northeast of the Thetford Mines area. MacKay (1921) and Tolman (1936) also show ultramafics in a basal volcanic member of the Beauceville in the same general region. In view of the evidence farther to the south, however, it may be well to question whether the beds mapped as basal Beauceville in the regions northeast of Thetford Mines may not actually belong to an older series. The writer tentatively dates the emplacement of the ultramafics as post-Cram Hill and pre-Shaw Mountain.

Granite and pegmatite

Small bodies of apparently intrusive rock ranging in composition from granite to quartz-diorite, some with associated pegmatites, are common in the basement rocks of the Green Mountain area and in both the basement and younger rocks in the vicinity of the Chester dome.

The granite and pegmatite in the Green Mountain basement rocks are clearly older than the Tyson formation. The Tyson conglomerates carry pebbles of the granite and large fragments of feldspar probably derived from the pegmatite. In at least one locality (Plate IV) a pod of pegmatite in the basement rocks is truncated by an arkosic conglomerate of the

Tyson formation. Some of the granite and pegmatite masses are of mappable size but have been omitted from the map for simplicity. The pegmatites and granites are similar, mineralogically, containing quartz, microcline, albite, muscovite and biotite. The microcline is commonly perthitic and pink in color. Black tourmaline is unusually abundant in many of the pegmatites. The pegmatite and granite appear to be genetically related inasmuch as gradational types between the two textural extremes are common. Cataclastic features are characteristic of the basement granites and pegmatites. Most of the granites are gneissoid and the pegmatites typically have a brecciated appearance. Some of the larger tourmaline crystals have apparently been broken and re-cemented at a later date by quartz, forming a mosaic. The plagioclase feldspars commonly show a coarse saussurite resembling that in the surrounding gneisses.

In the Chester dome area the granite problem is complicated by the fact that there is definite evidence of at least two generations of such rocks. The older granites and pegmatite intrude only the Reading gneiss, Whitesville marble and Cavendish schists, but the younger granites cut all formations including the Waits River. That the division of the granitic rocks of the Chester dome area into an older and a younger series is valid, is brought out rather strikingly in a small plug of the younger granitic rock in the southwestern part of Baltimore. The plug, measuring only a few hundred feet across, is surrounded entirely by Reading gneiss, but contains abundant large, disoriented xenoliths of rock types foreign to the immediate vicinity. The xenoliths include coarse albitic schists, garnetiferous schists, and Ca-Mg silicate rock, apparently derived from the Cavendish schist and Whitesville marble, as well as a few blocks of the surrounding Reading gneiss. The nearest outcrops of either the marble or schist are on Hawks Mountain 2,900 feet northwest of the xenoliths and 500 feet above. Projection of the marble and schist along the bedding planes, however, would carry them to within 1,300 feet, vertically, of the xenoliths. A remarkable feature of several of the larger schist xenoliths is that they are cross-cut by dikes of the older type aplite and granite which are in turn truncated by the younger granitic rock at the border of the xenolith. Besides showing the existence of two generations of granitic rock, the occurrence demonstrates effectively the once fluid nature of the younger granite. Tilted xenoliths, however, have been observed at other localities in granitic rocks of both groups.

The rocks of the older group contain as major constituents, microcline (perthitic and of a pinkish color in the pegmatite), albite, muscovite, and biotite. The minor constituents include epidote, or clinozoisite, and tourmaline. Epidote and clinozoisite, although not generally considered typical constituents of pegmatite, occur in conspicuous amounts in some of the larger pegmatites, particularly those at or near the horizon of the Whitesville marble. The epidote occurs as sprays of euhedral prismatic crystals, in some instances six or eight inches in length. A high calcium content of the pegmatites is not surprising in the vicinity of carbonate rocks but that the calcic mineral is epidote or clinozoisite rather than scapolite or calcic plagioclase is puzzling. Inasmuch as the occurrence in sprays of prismatic crystals suggests the mode of occurrence of scapolite at carbonate-pegmatite contacts in other localities (Bolton, Mass., for example) the writer interprets the epidote crystals as of secondary origin probably dating from the more recent regional metamorphism, the crystals probably occupying the sites of former scapolite crystals. As calcic scapolite plus water is identical in composition to clinozoisite plus carbon dioxide, the explanation is at least plausible.

All gradations apparently exist between aplite, granite and pegmatite, but there is a pronounced tendency for the extreme types, aplite and pegmatite, to occur together, the aplite commonly forming a border three to six inches wide along the sides of many of the smaller pegmatite dikes. In some instances, dikes of pegmatite cut aplite and in others the reverse is true, although both pegmatite and aplite appear to be always younger than the granites with which they are associated.

The older granites and pegmatites of the Chester dome are similar to those of the Green Mountain basement in that they characteristically have a gneissoid or brecciated appearance. It is possible that some of the clinozoisite in these granites, showing a tendency to be associated with the plagioclase, may be derived from the recrystallization of initially more calcic plagioclase.

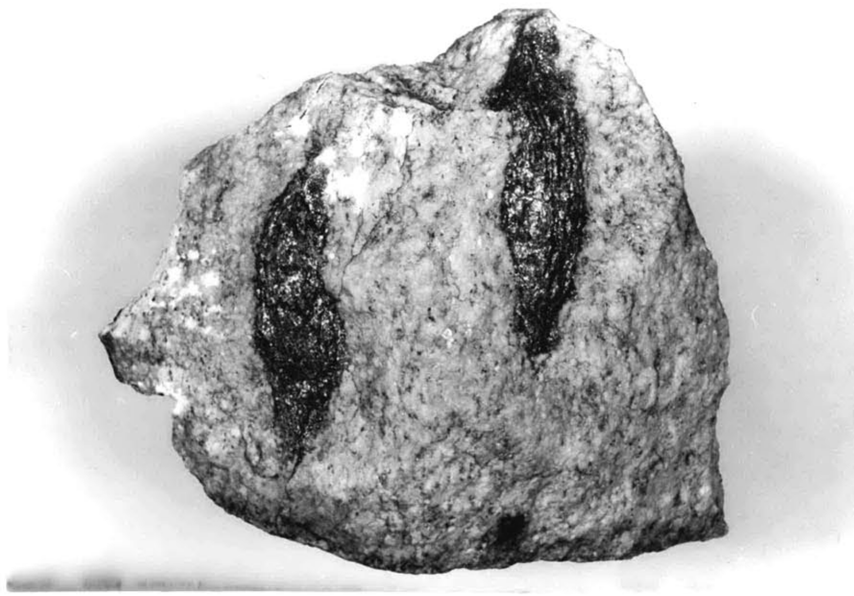
The younger granitic rocks of the Chester dome area differ from the older types in that microcline tends to be a subordinate constituent, or absent, and in the lack of cataclastic effects, although weakly foliate in some occurrences. In their lack of potash feldspar, however, these granitic rocks are more popularly quartz-diorites or granodiorites than granites. Maynard (1934) describes two of the younger granitic rocks of the Chester dome area as leucotonalites because of their low content of dark minerals. The plagioclase in the younger granites is albite or oligoclase. Other constituents are muscovite, biotite and clinozoisite. The clinozoisite, as in the older rocks, is associated with the plagioclase. Biotite is much less abundant than muscovite, some of the rocks containing almost no dark mineral. The only rock of the younger series containing biotite in large amounts is the orbicular quartz-diorite occurring in the Barnard gneiss just west of Proctorsville Village. The biotite in this locality, however, is concentrated almost entirely in the orbicules. The micas, both biotite and muscovite, in the

Plate VIII.

Orbicular quartz-diorite, Proctorsville

The flattened orbicules in this specimen are about three inches long.

Plate VIII



orbicules are concentrically arranged. The orbicules have the appearance (Plate VIII) of elongate and somewhat flattened prunes. The long axes of the orbicules are parallel to the fold axes in the surrounding Barnard gneiss and the flattening is roughly parallel to the foliation in the gneiss. Orbicular granitic rocks similar to the Proctorsville occurrence are found at several localities farther north in Vermont, notably Bethel, Craftsbury and Northfield, as well as at Stanstead, Quebec and to the south at Newfane, Vermont, near the southern end of the Chester dome. At the northern localities the orbicules tend to be approximately spherical in shape in contrast to the Proctorsville and Newfane occurrences in which they are ellipsoidal.

The writer has observed no aplites that could be definitely related to the younger granitic rocks. Pegmatite-like masses, however, do occur in some of the younger formations, but commonly as boudinage fillings of questionable genetic relationship to the granitic rocks.

The younger granitic rocks cut all formations including the Waits River. The slight foliation parallel to that in the surrounding schists and gneisses, however, and the elipsoidal nature of the orbicules seem to suggest that some, at least, of the deformation must have followed their emplacement. A similar relationship to the metamorphism is indicated by the occurrence of fairly large grains of clinozoisite, unlikely as a primary constituent of an igneous rock, but a possible product of its recrystallization at lower temperatures.

Veins and mineral pockets

Quartz veins and quartz lenses are abundant throughout the area mapped, occurring in all rocks except the ultramafics. In addition to the veins and lenses, quartz is of common occurrence in boudinage fillings, in which it is sometimes the only mineral present.

In the basement rocks of the Green Mountain area, many of the quartz veins contain abundant black tourmaline, some of the veins consisting almost entirely of tourmaline. Most of the tourmaline is in coarse crystals one eighth to one-half inch in diameter, but tiny needles are also common in some of the smaller veins. Lenticular or irregular pockets of the same composition are common in the same general area. Some of the pockets and veins also contain muscovite, albite, and pink, perthitic microcline and appear to be gradational into pegmatite.

In the younger rocks of the Green Mountain area and in the Chester dome rocks, the materials forming veins and boudinage-fillings show an interesting tendency to be similar in bulk composition to the rocks in which they occur. These include masses of quartz and zoisite in the Waits River formation; quartz, chlorite, albite and magnetite or ilmenite in the Pinney Hollow schists; and quartz, chlorite, magnetite and calcite in greenstones. Tourmaline has been observed in nearly all such occurrences. In feldspathic rocks such as the greywackes, arkoses, albite schits and their various more highly metamorphosed equivalents, the boudinage fillings are commonly feldspathic with a bulk composition and texture approaching that of pegmatite. In the Tyson formation of the Green Mountain area, material of this type forms boudinage-fillings and irregular veins and pockets, but is quite distinct from the pegmatite of the basement in that it is finer-grained, occurs in much smaller masses, and does not show the cataclastic and retrograde effects typical of the basement pegmatites..

Greenstone dikes and sills

The only mass of greenstone that is clearly intrusive into the surrounding rocks is the elongate body in the extreme southeastern corner of Plymouth. Although mapped by Perry (1928) as serpentine, this rock consists of actinolite, chlorite, calcite, ankerite, sodic plagioclase and epidote with what appears in the hand specimen to be a relictdiabasic texture. The body is non-foliate, and cuts across the schists of the Moretown formation at a low angle.

In the southeastern corner of Ludlow a similar body occurs in the greenstones of the Pinney Hollow formation. Unlike the body in Plymouth, it is concordant but has been interpreted, because of its coarseness of grain, as a sill rather than as a flow. It is about one half mile long and

100 feet thick. Its appearance in thin section is somewhat similar to that of the Plymouth occurrence but amphibole and epidote are relatively more abundant indicating a more mafic rock.

Both masses are clearly older than the metamorphism. Their mineral and chemical composition, moreover, is very similar to that of the nearby banded greenstones suggesting that they may be genetically related to them, and possibly not greatly different in age.

Quartz porphyry dikes and sills

Numerous small dikes and sills of quartz-albite-porphyry, generally less than five feet wide, have been observed cutting the basement rocks in the Tyson and Grahamville formations in the southern part of Plymouth. The albite phenocrysts are euhedral and contain abundant flakes of coarse sericite. The quartz phenocrysts are anhedral and free of inclusions. The ground mass consists of a relatively fine aggregate of quartz, sericite, albite, biotite and chlorite with traces of magnetite and epidote.

The porphyry dikes, like the greenstone dikes, have clearly been metamorphosed, the porphyritic texture indicating, furthermore, that the wall rocks were probably cool at the time of intrusion. In their high silica content, high alumina-alkali ratio and high soda-potash ratio, the dikes resemble both the supposed dacites of the Barnard gneiss and the small bodies of quartz diorite in the Chester dome area.

<u>Alkalic Stocks</u>

Two eruptive complexes, distinctly younger than the major deformation and metamorphism, outcrop within the area mapped. The larger and better known of these is the Mount Ascutney complex (Daly, 1903; Chapman and Chapman, 1940) which cuts across the northeastern margin of the Chester Dome. The smaller Cuttingsville stock (Eggleston, 1918) lies in the core of the Green Mountains southeast of Rutland. Syenite or "quartz-syenite" (syenite with less than 5% quartz) is the dominant type in both bodies, comprising nearly all of the Cuttingsville mass. The Ascutney complex also includes sizeable areas of gabbro, biotite-granite and a breccia of supposed volcanic origin. Alkalic amphiboles and pyroxenes are characteristic of most of the rock types. Fayalite is a common accessory, particularly in the quartz-syenites. The rocks are in general coarse-grained, fresh, massive and lacking in directional structures.

The stocks show no apparent relationship to any of the pre-existing structures and have not noticeably modified any of these structures, even in the immediate vicinity of the contacts. The persistence of the regional planar and linear structures in the vicinity of Mount Ascutney has been noted previously by Daly (1903) and by Chapman and Chapman (1940), and has been an important factor controlling their respective hypotheses of the mode of intrusion. Contact metamorphic effects are local and depend largely on the nature of the country rock.

The gneisses of the Green Mountain core and the Chester Dome are but little affected, whereas the phyllites and impure limestones of the Northfield and Waits River formations pass into a zone of massive, flinty hornfels which may be a hundred yards in width.

Petrographically the stocks have many features in common with other eruptive complexes in New England and eastern Canada, in particular with the ring-dike and related complexes of the White Mountain region of New Hampshire, the Cape Ann and Quincy batholiths of eastern Massachusetts and the Monteregian intrusives of southern Quebec. Their age is not definitely known (Williams and Billings, 1938) but can be bracketed with some assurance as either late Devonian or Missippian. The Vermont and Quebec bodies cut Middle and Upper Ordovician rocks, and, if the Mount St. Helens breccia is genetically related, Silurian as well. The New Hampshire bodies cut rocks of probable Lower-Middle Devonian age, and the Massachusetts representatives cut rocks of known Middle Cambrian age. The best evidence for an upper limit is in eastern Massachusetts where the Quincy granite is overlain unconformably by Pennsylvanian sediments.

Trap and Felsite Dikes

Fresh, fine-grained dike rocks showing a considerable range of composition have been found over nearly all of the area mapped. The greatest concentration, however, particularly of the more felsitic types, is in the vicinity of the Ascutney and

Cuttingsville stocks. In the region west of Mt. Ascutney, moreover, there is a definite tendency to radial arrangement about the stock. The writer has done no petrographic work on the dikes but from their distribution in relation to the alkalic stocks it is probable that the bulk of them are genetically related.to the stocks. Farther south in the Connecticut Valley trap dikes clearly related to the Triassic igneous activity are common. It is likely that some of the trap dikes in the area mapped are Triassic but impossible without detailed petrographic work, and perhaps not even then, to say which.

The felsite dikes are commonly finer-grained than the traps, and most of them contain no visible quartz. According to Balk (1936) some of the felsites near Ascutney are devitrified glass. Some of the darker dikes, notably on the west slopes of Blueberry Mountain in Plymouth, are porphyritic with phenocrysts of black hornblende reaching an inch in diameter. The dikes vary in thickness from a few inches to about one hundred fifty feet. Most of the larger dikes are felsite, the trap dikes rarely exceeding six or eight feet in thickness.

STRUCTURE

Major Structures

The Green Mountain anticlorium

The Green Mountain anticlinorium can be traced from a point just south of the Vermont-Massachusetts line, near Williamstown, Massachusetts, northward at least to the latitude of Middlebury. Whether the anticlinorium of the northern Green Mountains and the Sutton Mountains of Quebec is a continuation of this same axis, or forms another slightly offset to the east, is uncertain at the present time.

Between Wallingford and Clarendon on the western side of the range, the rocks of the Mendon series and Chester quartzite form little more than a veneer, dipping steeply off the western front of the old crystalline massif and beneath the carbonate rocks of the Vermont Valley. Dips along the mountain front average between fifty degrees west and vertical, flattening westward toward the center of the valley, which at least between Rutland and Danby, is synclinal. Between Bear Mountain, in Wallingford, and Clarendon Gorge, however, the strata are overturned with easterly dips as low as sixty degrees. No evidence of large-scale thrust-faulting has been found by the writer, despite frequent references in the literature to the "Green Mountain border fault". Apparently the rather imposing scarp along the western front of the Green Mountains is the result of the removal by erosion of the relatively non-resistent carbonate rocks of the valley from the surface of the steeply dipping Cheshire quartzite. The anticline west of the Rutland Valley, however, is broken by the Pine Hill thrust (Dale, 1894; Fowler, 1949), and the Cheshire quartzite and Mendon series are thrust westward onto Ordivician strata of the Middlebury synclinorium (Cady, 1945). On Fowler's cross section (1949) the fault plane is shown as having an easterly dip of about forty degrees.

In the Ludlow area, on the eastern flank of the anticlinorium, the conglomerates and greywackes of the Tyson formation dip between forty and fifty degrees to the east. Farther east, however, toward the axis of the Proctorsville syncline, the dips steepen progressively and are vertical in the vicinity of the synclinal axis. Subsidiary folds, large enough to be mapped, are almost non-existent on the eastern limb of the Green Mountain anticlinorium in the Ludlow area.

Folds too small to map, however, are abundant throughout the area, plunging gently northward and showing the normal shear-sense for the eastern limb of an anticline (Plate IX-a). The gentle plunges of these small folds are probably the reason for the remarkable straightness of the formation-boundaries and their parallelism with the strikes.

Several high-angle thrusts offset the contact between the Tyson formation and the basement rocks in the southern part of Plymouth. As nearly as could be determined, the

Plate IX.

Minor folds, Chester and Ludlow.

(A)

Drag fold, west side of Chester dome.

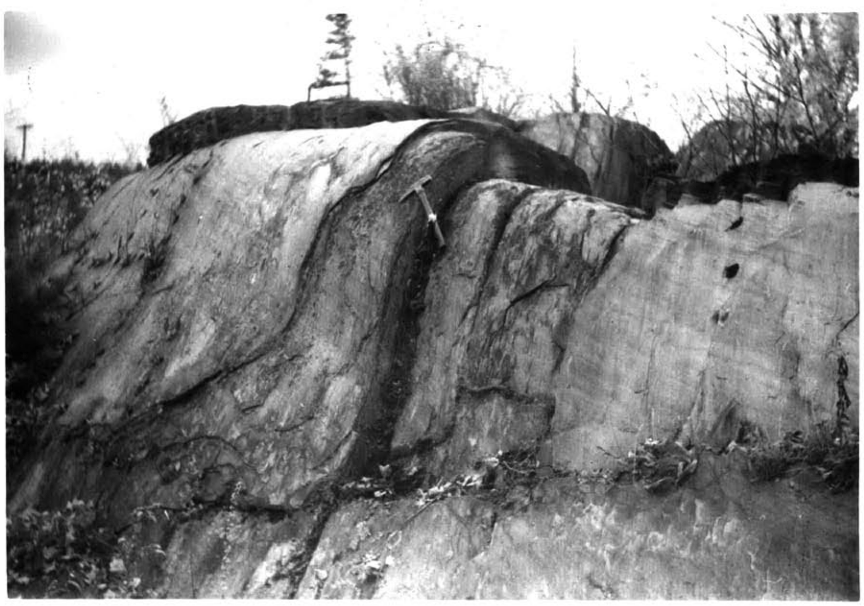
The locality is about 2 miles northwest of Chester village. The camera was pointed north about parallel to the plunge of the fold. The rock is banded epidote-amphibolite of the Pinney Hollow formation. The pits are from the weathering of calcite.

(B)

Minor fold, Green Mountain anticlinorium.

The locality is just west of the Ludlow village limits. The fold is in thinly bedded, impure quartzites of the Plymouth member of the Grahamville formation. The "plaid" effect on the dip-slope nearest the camera is formed by the intersection of crinkling (parallel to the fold axis) and streaming (approximately normal to the fold axis).

A



B

fault planes dip between fifty and fifty-five degrees to the east. It is interesting that this zone of faulting coincides with the region where the Ludlow Mountain schists and quartzites intersect the unconformity with the younger rocks. Apparently, appropriately oriented schistose bands in the predominantly gneissic basement complex were the loci of strong differential movements during the Paleozoic orogeny.

The trend of the Ludlow Mountain band of schists, however, is not typical of the basement trends in general. The writer made no serious attempt to unravel the Green Mountain basic structures, but reconnaissance in the Wallingford quadrangle, and in the southern part of the Rutland quadrangle, indicates that the prevailing strikes in the central part of the massif vary from east-west to northwest-southeast, with variable, but for the most part gentle, northerly and northeasterly dips. Toward the eastern side of the massif the strikes tend to swing around in a clockwise manner and the dips to steepen so that the attitudes in the older and younger rocks are commonly sub-parallel in the vicinity of the unconformity. This tendency is well displayed by the schist-quartzite bands forming the ridges of Markham Mountain and the southern peak of Terrible Mountain in Andover. Toward the western side of the massif, on the other hand, strikes swing around in a counter-clock-wise manner, with steepening dips, so that the basement strata are commonly sub-parallel to the younger strata as on the east. This deflection of the basement trends is consistent with their later arching about a fold axis plunging gently to the north. It is significant that the schist-quartzite band forming the northernmost ridge of Terrible Mountain does not swing around in this manner and has a west-northwesterly strike and almost vertical dip. The bedding planes apparently are normal to the younger fold-axis and therefore not deflected.

Although the younger strata have been removed by erosion from the central portion of the Green Mountain anticlinorium, it is clear from the dips on the flanks that the fold as a whole is asymmetric with a tendency to be overturned toward the west, suggesting a slight counter-clockwise (as viewed on a standard east-west cross section) rotational sense to the movements producing the deformation. Inasmuch as all of the known fault planes in the vicinity of the anticlinorium dip easterly with movements in the same counter-clockwise rotation sense, it is probably safe to assume that the deformation of the Green Mountain area may be related to a large scale couple involving upward and westward movement of the area east of the Green Mountains relative to that to the west.

In its overall dimensions, subsidiary features, and general relationship to the orogenic belt, roughly marking the boundary between the zone of regional metamorphism and the zone of simple folding and over-thrusting, the Green Mountain anticlinorium is similar to the other great anticlinoria more or less on strike with it to the south. These are the Berkshire, Housatonic Highlands, and Jersey-Hudson Highlands anticlinoria, and the Blue Ridge anticlinorium of the southern Appalachians. The central massifs of the Alps, in particular the Aar and Mont Blanc massifs, also appear to have many fundamental features in common with the Green Mountain anticlinorium.

<u>The Chester dome</u>

The Chester dome is about thirty-seven miles long and between six and eight miles wide, differing from most similar structures elsewhere in New England in being somewhat larger and more elongate. The banding in the central gneisses is horizontal or sub-horizontal over much of the area of the dome, with the dips steepening rather suddenly, not only along the sides of the dome, but at the ends as well. Subsidiary folds, many of them of mappable size, are common along the sides of the dome but are remarkable in that they indicate rotational movements precisely the reverse of those on the flanks of a normal anticline (Plate IX-B). In the areas mapped by the writer, most of these folds have gently northerly plunges, although plunges are southerly toward the southern end of the dome, in the Saxtons River quadrangle. The large subsidiary fold along the western side of the dome is rather complex and has been named, for convenience in discussion, the Butternut Hill fold. Its pattern, however, like that of the smaller folds, is that of a reverse drag.

Another remarkable feature of the dome is the fact that the formations surrounding it are thinner than the same formations in the Green Mountain area by a factor of approximately ten. Boudinage, flattened pebbles and other features indicate that the thinning is probably tectonic, the rocks mantling the dome apparently having been stretched like the shell of a blister.

The Proctorsville syncline and its relations to the Butternut Hill fold provide the most complex structural problem in the area. On the western side of Proctorsville Gulf where the band of the Northfield and Waits River formations in the Proctorsville syncline reaches its southern limit, it is not possible to determine whether the base of the formation actually "goes around end", or whether the formation has been pinched out of the axial-plane region of the fold in a sort of large-scale boudinage. In the north central part of the Saxtons River quadrangle, however, (Plate I) a small tear-drop shaped area of the Shaw Mountain, Northfield and Waits River formations occurs, forming the Spring Hill syncline as mapped by Rosenfeld (personal communication). At its, southern end the Spring Hill syncline appears to be a normal, northward-plunging syncline with the Shaw Mountain conglomerates "going around end" in the approved manner. At its north end, however, the relationships are almost a mirror image of those at Proctorsville Gulf. The minor folds, however, plunge north and are thus inconsistent with interpretation of the Spring Hill fold as a doubly plunging syncline. The evidence of the minor folds was, nevertheless, at first ignored inasmuch as the Spring Hill fold appeared to be offset too far to the east of the Proctorsville syncline to have any possible connection with it, and appeared to be a separate syncline separating the gneisses of the Butternut Hill fold from those of the Chester dome proper. Later field work, however, showed that the Pinney Hollow formation bordering the gneisses of the Butternut Hill fold dipped northward at the southern end of the fold beneath the gneisses, and that nowhere in the supposed northern half of the Spring Hill syncline did the minor folds show the expected reversal to a southerly plunge. Further evidence on the problem came with the discovery that the Moretown-Cram Hill contact on the western limb of the Proctorsville syncline also curved around the southern end of the Butternut Hill fold, indicating that the axial plane of the Proctorsville syncline must do the same. The present interpretation, shared by Rosenfeld, is that the Proctorsville and Spring Hill synclines are part of the same large isoclinal fold, the axial plane of which has been offset by the Butternut Hill fold, which has the form of a large-scale reverse drag fold on the western side of the dome. The Spring Hill syncline is thus interpreted as the detached "keel" of the Shaw Mountain-Northfield-Waits River sequence in the Proctorsville syncline.

The only key to the older basement structures in the Chester dome is the Hawks Mountain complex. This appears to be a nearly isoclinal synclinorium, overturned to the south, with its axial plane arched by the later doming. The east-west trend and the northerly dips are in essential agreement with the prevailing trends in the Green Mountain basement complex. The modification of the older structures, however, toward the margins of the dome is much more intense than on the limbs of the Green Mountain anticlinorium and probably is an important factor in the obliteration of the unconformity.

The reversal of shear sense on the sides of the dome and the thinning of the overlying formations over the crest are precisely the reverse of the conditions to be expected in a typical fold formed by lateral compression, and are inconsistent with the mechanics of such a fold. Vertical upward movement of the central mass with stretching of the overlying rocks, however, is quite consistent with the observed relationships. The Chester dome seems to have most of the characteristcs of a "mantled gneiss-dome" as defined by Eskola (1949).

MINOR STRUCTURES AND MICRO-STRUCTURES

Minor folds

Folds, too small to be shown on the geologic map, have been classified as minor folds, the classification thus including everything from folds many yards across to microscopic crinkles of the foliation. Most such folds are asymmetric and may properly be called drag folds.

The shear sense of the drags is persistent over large areas and consistent with the larger scale structural pattern. An interesting feature, however, is the extreme contortion upon the short or drag limbs of many of the smaller asymmetric folds of the Chester dome area. The deformation in such instances may be quite spectacular, the convolutions showing no systematic shear sense or parallelism of the axial planes. The plunges of the fold axes, however, agree quite closely. Other striking effects occur in folds in rocks of greatly differing competency (Plate X), the more plastic material being squeezed into irregular shapes in the crests and troughs of the folds.

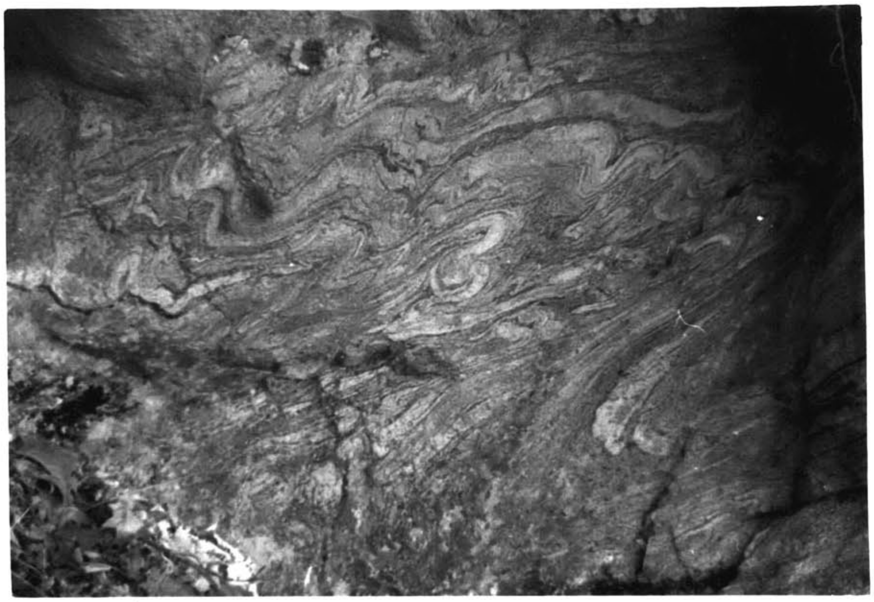
The minor folds may be either open or isoclinal, but open S-shaped folds are more generally the rule. The wave length and amplitude of minor folding seems to be in direct proportion to the competency of the rock. In the highly micaceous schists folds more than a few inches in wave length are rare, most of the crinkles averaging about a centimeter from crest to crest. In the quartzites and gneisses, on the

Plate X

Minor folds, Butternut Hill area

The locality is near the summit of the conical hill (el. 1520') about 1½ miles SSW of Butternut Hill in Chester. The rock is the thinly-banded phase of the Reading gneiss. The folds in (B) provide a good example of the squeezing of incompetent, schistose material into the crests and troughs of folds.

Plate X



A

B

other hand, the folds are generally several feet from crest to crest.

It is clear from the tectonic map (Plate III) that the plunges of the axes of the minor folds are persistent, areally, and for the most part are gently to the north. The most notable exception occurs just north of the block of gneiss forming the summit of Dry Hill. Here the minor folds over a small area plunge nearly down the dip of the foliation and are thus nearly at right angles to the prevailing fold plunges. It is noteworthy, however, that these downdip folds show no consistent asymmetry or shear sense.

The minor folds in the basement rocks of the Green Mountain massif were not studied in detail but tend to have easterly or southeasterly plunges.

Foliation

The pronounced foliation shown by the majority of rocks studied results, in general, from the parallelism of platy minerals, but may also result from a random orientation of rod-like minerals in parallel planes. The foliation in the compositionally banded rocks is almost invariably parallel to the banding. This is almost rigorously true in the younger stratified rocks in the Green Mountain area and true to a somewhat lesser extent in the rocks of the Chester dome. The only exception observed in the younger rocks of the Green Mountain area is in the faulted zone on Dry Hill in the southern part of Plymouth. On the south side of the isolated block of gneiss forming the summit of the hill, the foliation cuts sharply across the bedding of the conglomerate and across the plane of unconformity between the conglomerate and the older gneiss*. Elsewhere in the Green Mountain area foliation and banding, wherever both can be identified, are, within the limits of the writer's observation, strictly parallel, even around the noses of sharp, sometimes isoclinal, folds. Inasmuch as primary stratification could be determined with certainty in more instances than not, the evidence for parallelism must be considered as largely positive. There is no evidence whatsoever of an earlier generation of folds with axial planes parallel to the foliation. Slip-cleavage, to be discussed more fully in a later section, does, however, form an approximate axial plane cleavage in some of the younger folds.

The discordant foliation occasionally observed in the Chester dome rocks occurs chiefly on the noses of small drag folds. The minerals forming the discordant foliation, however, are commonly in bands separated by regions in which the micaceous minerals are parallel to the banding. In such occurrences the writer has interpreted the discordant foliation as a re-crystallized slip-cleavage.

The best examples of discordant foliation were observed in the basement rocks of the Green Mountain massif, in the vicinity of their contact with the overlying strata.

*See p. 104.

Foliation of this sort is approximately parallel to the plane of unconformity and hence to the bedding and foliation in the younger rocks above.

Some of the younger unstratified rocks show distinct foliation. It is particularly pronounced in the serpentines and associated talc-carbonate rocks, but has also been observed in some of the younger granitic rocks although never as well developed. The foliation in the unstratified rocks is approximately parallel to that in the surrounding schists and gneisses in most instances.

Streaming and Mineral Lineation

The word "streaming" has been used in the field to describe a streaking-out of mineral aggregates in the plane of foliation. The streaming may or may not be accompanied by a dimensional orientation of the mineral grains parallel to the long axis of the aggregate. It is apparently similar to the "Striemung" of European geologists. Cloos (1940) translates the German as "streaming". F. C. Phillips (1947) has described a lineation in the Scottish Highlands which appears to be quite similar to that in Vermont.

The streaming is best developed in the Green Mountain region where it is characteristically oriented very nearly down the dip of the foliation and approximately normal to the axes of the minor folds and crinkles. Deviations from normality between the streaming and the fold axes rarely exceed twenty degrees. A typical relationship is for the streaming to be almost precisely down dip, with the fold axis plunging about ten degrees to the north. The streaming shows up in most instances as elongate aggregates of chlorite or biotite flakes. The individual flakes generally do not show dimensional orientation parallel to the long axes of the aggregate. In some areas, however, the biotite occurs as small rhombic plates with their intermediate axes and cleavages normal to the foliation. The long axes of such plates are commonly parallel to the streaming when present, as are those of the small amphibole needles in the greenstones.

Eastward from the Green Mountains toward the axis of the Proctorsville syncline, the typical streaming becomes steadily less conspicuous, giving way to a simple mineral lineation of amphibole needles or biotite rhombs, which may be either normal (as in the typical streaming farther west), or parallel to the fold axis. In the vicinity of the synclinal axis and about the Chester dome the mineral lineations are typically parallel to the fold axes, although several examples of mineral lineation normal to the fold axes have been observed by the writer along the eastern side of the Chester dome. It is interesting that west of the axis of the Proctorsville syncline, in the transition zone between mineral lineations predominantly normal to the fold axes and those predominantly parallel to the fold axes, it is not uncommon to find lineations of both types in the same outcrop. Exposures have even been observed where approximately half the amphibole crystals in one foliation plane are normal to the fold axis and the others parallel to it, the two sets of amphibole crystals apparently identical in every way but orientation. It is also not uncommon to find, in this transition zone, and in many regions of the Chester dome as well, that the elongate minerals have a random orientation in the foliation plane.

Slip-cleavage

Although the foliation through the Green Mountain Region and much of the Chester dome area is strictly parallel to the bedding even in the crests and troughs of folds, it is not uncommon to find a cleavage, distinct from the foliation, which is decidedly discordant to the bedding. Following Dale (1896) this phenomenon has been called slip-cleavage. Slip-cleavage results from the parallelism of platy minerals along the limbs of small crinkles in the schistosity. All stages in the development of slip-cleavage may be seen, sometimes within the confines of one thin-section. Where the folds are asymmetric the development of slip-cleavage normally ends in complete shearing out of the short or drag limb of the crinkle (Plate XI). If the folding is symmetrical, however, shearing commonly does not take place, and the crinkles form small similar folds in which the cleavage results from the parallelism of the micas on their limbs. Where the crinkles appear as small drags upon the limbs of a larger fold, the

Plate XI.

Slip-cleavage.

The specimen in (A) is from the base of the Pinney Hollow formation at the westward bend in Buffalo Brook, Plymouth.. The specimen in (B) is from the upper part of the Grahamville formation east of Lake Amherst in Plymouth.

A

B

attendant slip-cleavage planes are approximately parallel to the axial planes of the larger fold.

Some of the best displays of slip-cleavage occur in the banded "pin-striped" quartzites typical of the Moretown formation. This slip-cleavage apparently results from the extreme thinning of the granoblastic laminae on the limbs of small similar folds.

The slip-cleavage west of the Plymouth-Ludlow valley is, on the average, either vertical or dipping steeply to the east. Farther east, westerly dips prevail, becoming increasingly gentle in the direction of the Chester dome. Good examples of slip-cleavage are rare in the dome area. The observed occurrences, however, are generally parallel to the axial planes of the minor folds, suggesting that the slip-cleavage, if better developed, would form an arch across the dome, somewhat flatter than that of the bedding.

Slip-cleavage is a familiar feature to most geologists who have worked in the crystalline rocks of western New England. The most complete descriptions are those of T. N. Dale (1896, 1914), H. E. Hawkes (1941), and W. S. White (1949). Although each of the above use the term "slip-cleavage", first proposed by Dale, the phenomenon has gone by other names such as "fault-slip-cleavage" (Leith, 1905) and "shear-cleavage" (Mead, 1940). Leith, noting that actual shearing of the limbs was not an essential feature of the cleavage, considered slip-cleavage to be essentially a variety of fracture-cleavage, but different from true fracture-cleavage in that actual displacement along the cleavage planes need not have taken place. In the Swiss Alps similar cleavage phenomena have been called "Ausweichungs-clivage" (Heim, 1878, pp. 43-4). The writer prefers "slip-cleavage" to the other English terms, particularly to "fracture-cleavage" which refers to a cleavage probably formed by a somewhat different mechanism in non-foliate rocks.

Rotation and deformation of porphyroblasts

Porphyroblasts with spirally arranged inclusions, indicating rotation of the porphyroblasts during growth, have been observed in many of the thin-sections. The phenomenon is most conspicuous in garnets but has also been observed in plagioclase augen and in amphiboles. The spirals are visible only in sections cut normal, or nearly so, to the fold axes, indicating that the fold axes probably coincide with the axes of rotation of the porphyroblasts. The direction of rotation, furthermore, is in agreement with the shear sense indicated by the minor folds and other features.

The amount of rotation is generally less than 180° (Plate XII-A), but in a few instances (Plate XII-B) is considerably more. Where the rotation is less than 180°, it might be argued that the spiral does not necessarily indicate rotation during growth but is merely the trace of a small crinkle or drag fold which served as the locus for the later development of the porphyroblast. If this were so,

Plate XII.

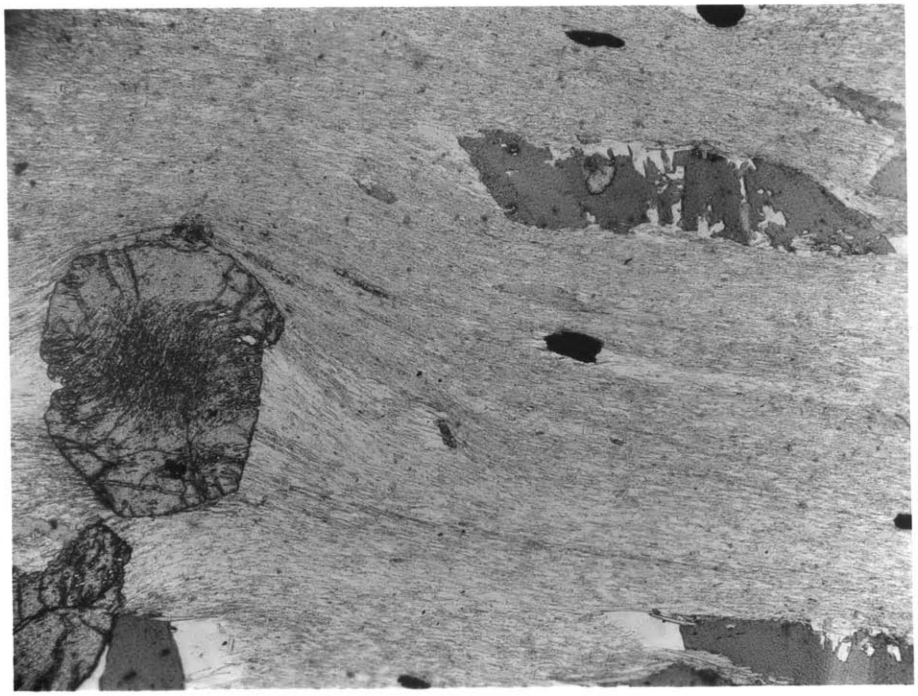
Totated porphyroblasts.

(A)

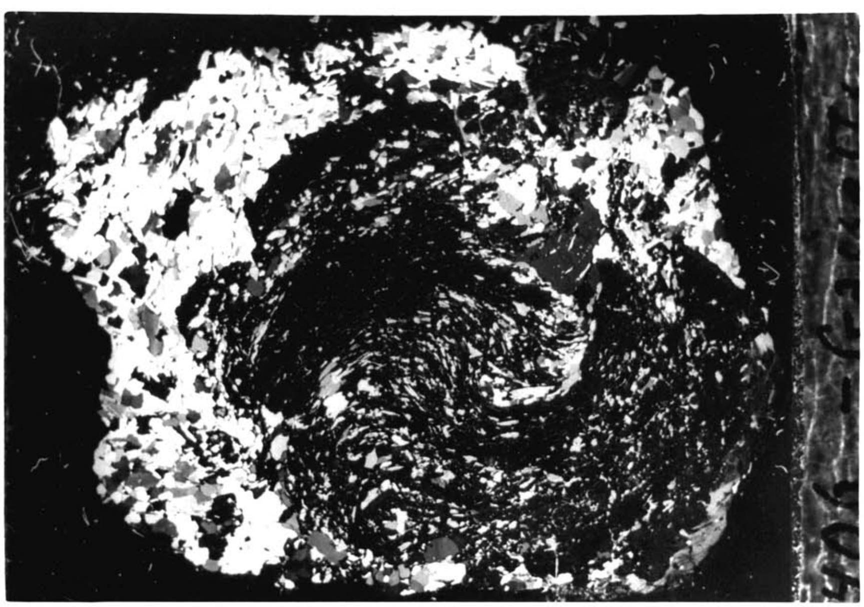
Rotated garnet and sheared books of biotite, Whetstone Hill member of Moretown formation, East Hill, Andover.

(B)

"Snowball" garnet from Gassetts schist near Whitesville.



A



B

however, the crinkle should be expected to extend along its axis beyond the limits of the porphyroblast, a feature which has not been observed.

The garnets of the Pinney Hollow and Stowe formatins in the Green Mountain region are commonly flattened parallel to a dodecahedral face. These tabular garnets may lie in the foliation plane but are more commonly rotated 45° or more from the foliation, about an axis parallel to the crinkle or fold axis and normal to the streaming. The rotation sense is clearly shown by the curvature of the schistosity in the vicinity of the tablet. The inclusions within the tablet generally show a slight spiral with the same sense of rotation, indicating that the rotation of the garnet was, at least in part, contemporaneous with its growth.

Rotational movement is also shown by deformed books of biotite (Plate XII-A), or chlorite, but unless the books are actually broken or faulted by the rotation, it is somewhat uncertain whether the movement was contemporaneous with, or later than, the growth of the crystal. Bending of mica flakes about small crinkles is quite common although in some instances the micas have crystallized or recrystallized after the folding to form polygonal arches similar to those described by Weiss (1949) in the Wissahickon schist of Pennsylvania. Bent crystals of kyanite are fairly common, particularly in some of the Cavendish schists on the northern slopes of Hawks Mountain. Some of these curved kyanite blades are in parallel growth with staurolite which, being more brittle, has ruptured. A thin-section of the sheared gneiss just east of the main thrust fault on Dry Hill in Plymouth shows plagioclase with bent twin lamellae.

Pebble deformation

The occurrence of conglomeratic rocks at several widely separated points has afforded an interesting opportunity to compare the deformation of pebbles in different tectonic environments. The conglomerates of the Tyson formation, west of the Plymouth-Ludlow Valley, and those at the base of the Mendon series in Wallingford are probably the most striking and the best exposed. These localities have considerable historical interest in that they are among the first occurrences of deformed pebbles to be described, and certainly the first to be discussed at any length, in American geologic literature (Hitchcock, 1861, pp. 28-45).

The Tyson conglomerates on Dry Hill in Plymouth dip between 40 and 50 degrees easterly and are offset by high-angle thrust faults also dipping to the east. The dips of the fault planes appear to be nearly the same as those of the conglomerate beds but must average slightly steeper inasmuch as older rocks (basement gneiss and Tyson conglomerate) are thrust over younger (schists of Grahamville formation). The pebbles, largely of quartzite, gneiss, vein quartz, and pegmatites, are flattened in the plane of foliation and are elongate down the dip of the foliation, parallel to the streaming. Over most of the hill the foliation is parallel to the bedding, even around the noses of several isoclinal drag folds measuring between five and fifteen feet across. On the south side of the hill, however, along the southeastern border of the isolated block of older gneiss, the bedding dips steeply westward, beneath the gneiss, whereas the foliation, streaming, and long axes of the pebbles dip eastward as elsewhere on the hill. Except at the locality previously described*, at the north end of Dry Hill, the folds plunge from ten to fifteen degrees north and have a shear sense indicating relative upward movement on the east. Where the foliation wraps around the noses of these folds the pebbles and streaming follow suit (Plate XIII). Plate XIV-B shows a porpoise-shaped pebble formed in this manner. The average axial ratios of the quartzite and gneiss pebbles vary from about 1: $1\frac{1}{2}$: 2 in the wedge-shaped mass of conglomerate south of the gneiss block to about 1 : 4 : 30 in the immediate vicinity of some of the faults on the eastern slope of the hill. The ratios in most of the outcrops, however, and in other areas of the Tyson conglomerate, average between 1 : 2 : 4 and 1 : 3 : 9. An interesting and inexplicable feature of many of the pebbles in the Dry Hill area is their tendency to have one rounded end and one pointed end, the rounded end, in most instances, being the upper (Plate XIV-A).

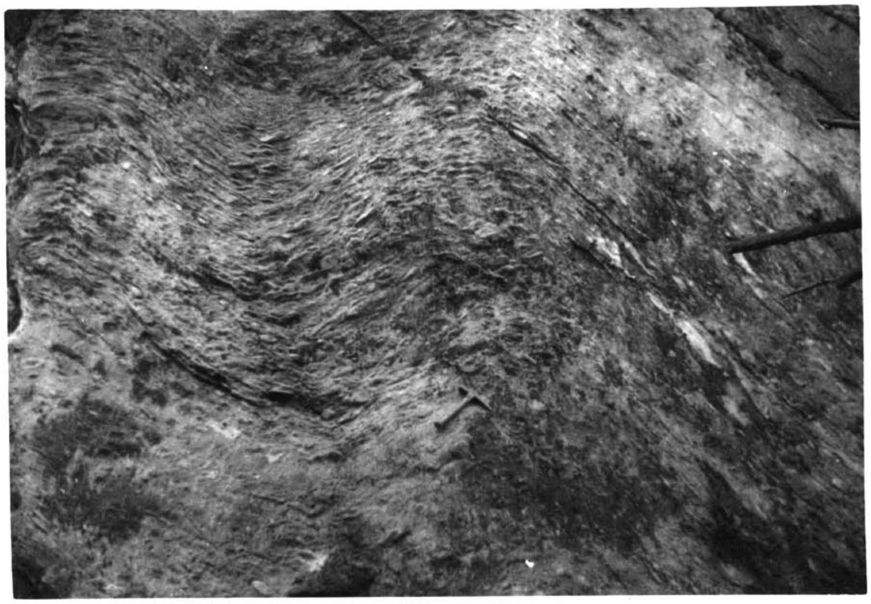
*See p. 95.

Plate XIII.

Folded conglomerate, Plymouth.

The locality is at an elevation of about 1340' on the southeastern slopes of Dry Hill in Plymouth. Both views are of the same fold, that in (B) approximately in the direction of plunge. The fold plunges about fifteen degrees almost due north. The slip-cleavage, which apparently provided access for the quartz vein, dips about sixty degrees east and is parallel to the axial plane of the Fold.

The pebble elongation is approximately normal to the fold axis.



A

B

Comparison of pebbles of different lithologies is revealing inasmuch as coarseness of grain rather than composition seems to be the controlling factor. Pebbles of quartzite and gneiss, of comparable coarseness of grain, appear to differ little if at all in their deformational characteristics, but pebbles of vein quartz, coarse pegmatite or feldspar probably derived from pegmatite often retain what must be very nearly their original shapes even in the zones of greatest deformation, showing only minor granulation about the edges and occasional shearing off of angular projections. At one point a nearly equidimensional pebble of feldspar, about one inch in diameter, was observed to indent sharply an adjacent quartzite pebble with axial ratios of about 1 : 3 : 9.

The conglomerates at the base of the Mendon series in Wallingford and Clarendon are similar to the Tyson formation in their deformation. The pebbles are flattened in the plane of foliation, again generally parallel to the bedding, and are elongate parallel to the streaming which is generally down the dip. The axial ratios, however, average about 1 : $1\frac{1}{2}$: 3 indicating less intense deformation than on the east side of the anticlinorium.

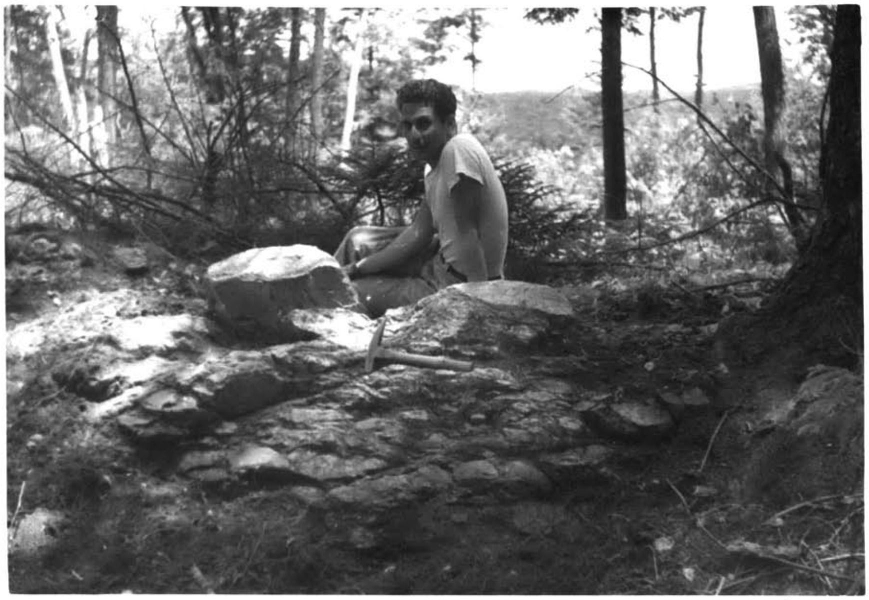
The pebbles in the Shaw Mountain conglomerates just west of the axis of the Proctorsville syncline along the ridge called "The Alps" are deformed in a manner quite different

Plate XIV.

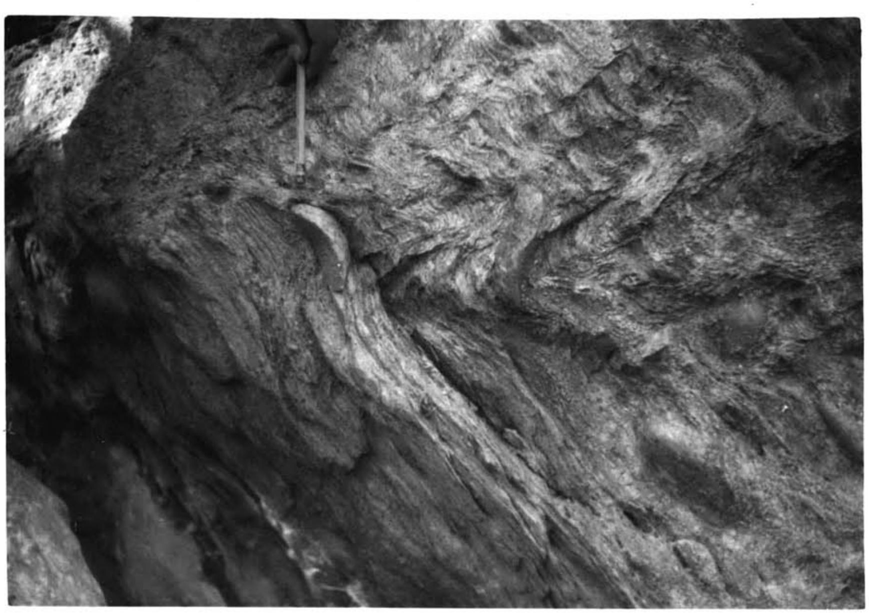
Deformed pebbles, Plymouth.

The locality of (A) is near that of Plate VI. The large pebble measures 8 x 24" in cross-section and has been partly isolated by removal of the schistose matrix. The locaity of (B) is just south of the Dry Hill gneiss-block. The slip-cleavage, dips about forty degrees west at this locality, but dips steeply east elsewhere in the Dry Hill area. Note rounded upper end and sharp lower end of pebble below pencil.

The elongation and fluting of the pebbles is normal to the fold axes and parallel to the streaming in the matrix.



A



B

from those of the Green Mountain area. Here the minor folds and crinkles plunge gently north and the pebbles are approximately cigar-shaped with elongation parallel to the fold axes, and a very slight flattening parallel to the foliation. The axial ratio (omitting the slight flattening) varies from 1 : 3 to 1 : 6. As nearly all of the pebbles are of finely granular quartzite it was not possible to compare pebbles of different lithology at these localities.

The only good conglomerate observed by the writer in the Chester dome area is in the Shaw Mountain formation in the southwestern part of Springfield township. The quartzite pebbles, like those in "The Alps", are elongate in the direction of the fold axes, here plunging 30°N 35° E, but are also flattened in the plane of foliation with axial ratios averaging about 1 : 6 : 30. The situation is further complicated in that some of the pebbles are themselves folded, producing rather bizarre cross sections. Rosenfeld (personal communication) has found pebbles deformed in a still different manner, at a horizon somewhat deeper in the dome, in a conglomerate believed equivalent to part of the Tyson formation. The pebbles are flattened to an extreme degree in the foliation plane but are nearly disc-shaped with a slight tendency to elongation normal to the fold axes.

Orbicule and quartz-pod elongation

The elongate nature of the biotite nodules in the granite at Proctorsville has already been mentioned. The axial ratio of the orbicules, omitting the slight flattening parallel to the foliation, varies from 1 : 3 to 1 : 4, nearly the same as that in the pebbles of the Shaw Mountain conglomerate a short distance to the north. Orbicules from granites elsewhere in Vermont are generally nearly spherical, except in the occurrence at Newfane near the southern end of the Chester dome. This suggests that the shape of the orbicules at Proctorsville is the result of deformation either following or during the emplacement of the granite.

The paddle-shaped quartz pods, characteristic of the Pinney Hollow and Stowe formations, are flattened in the foliation plane and are elongate down the dip of the foliation in the Green Mountain region, and parallel to the fold axes in the vicinity of the Proctorsville syncline.

Joints

Although no systematic study of joints was made, conspicuous joint systems, apparently related to the other structural features, were observed over much of the area. Joints normal or sub-normal to the fold axes are characteristic of the entire area but are most conspicuous in the vicinity of the Proctorsville syncline and the Butternut

Orbicule and quartz-pod elongation

The elongate nature of the biotite nodules in the granite at Proctorsville has already been mentioned. The axial ratio of the orbicules, omitting the slight flattening parallel to the foliation, varies from 1 : 3 to 1 : 4, nearly the same as that in the pebbles of the Shaw Mountain conglomerate a short distance to the north. Orbicules from granites elsewhere in Vermont are generally nearly spherical, except in the occurrence at Newfane near the southern end of the Chester dome. This suggests that the shape of the orbicules at Proctorsville is the result of deformation either following or during the emplacement of the granite.

The paddle-shaped quartz pods, characteristic of the Pinney Hollow and Stowe formations, are flattened in the foliation plane and are elongate down the dip of the foliation in the Green Mountain region, and parallel to the fold axes in the vicinity of the Proctorsville syncline.

Joints

Although no systematic study of joints was made, conspicuous joint systems, apparently related to the other structural features, were observed over much of the area. Joints normal or sub-normal to the fold axes are characteristic of the entire area but are most conspicuous in the vicinity of the Proctorsville syncline and the Butternut

Hill fold. The jointing is better developed, as a rule, in the more gneissic rocks than in the schists, presumably because of the relatively greater competency of the gneisses. In the Barnard gneiss on the western limb of the Proctorsville syncline and in the Reading gneiss of the Butternut Hill fold the jointing has apparently had considerable effect on the development of the topography as evidenced by the conspicuous east-west trending ridges in these areas. It is interesting that these ridges, although well developed in the gneisses immediately east and west, do not extend across the narrow band of schists defining the axial plane of the Proctorsville syncline. Some of the best bedrock exposures in the entire area occur on the south slopes of these joint ridges where glacial plucking was apparently facilitated by the steeply dipping joint faces.

A second well-defined set of joints is oriented, in general, roughly normal to the compositional banding or bedding and parallel to the fold axes. These joints, like the other set, are best developed in the relatively competent rocks such as quartzites and gneisses.

Boudinage

Boudinage, or "sausage structure", has been observed throughout the area mapped but is most common in the rocks of relatively higher grade metamorphism in the Chester dome (Plate XV), and in the basement complex of the Green

Plate XV.

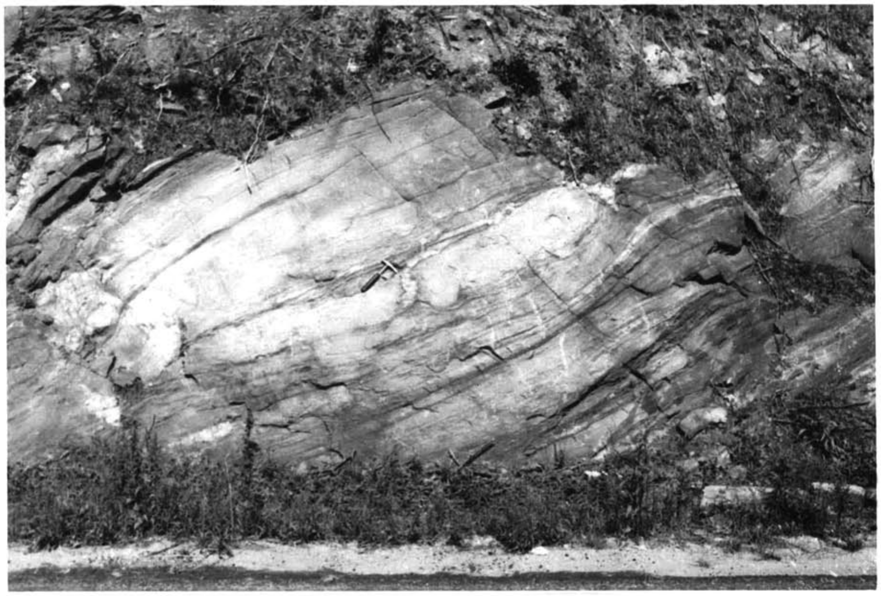
Boudinage.

(A)

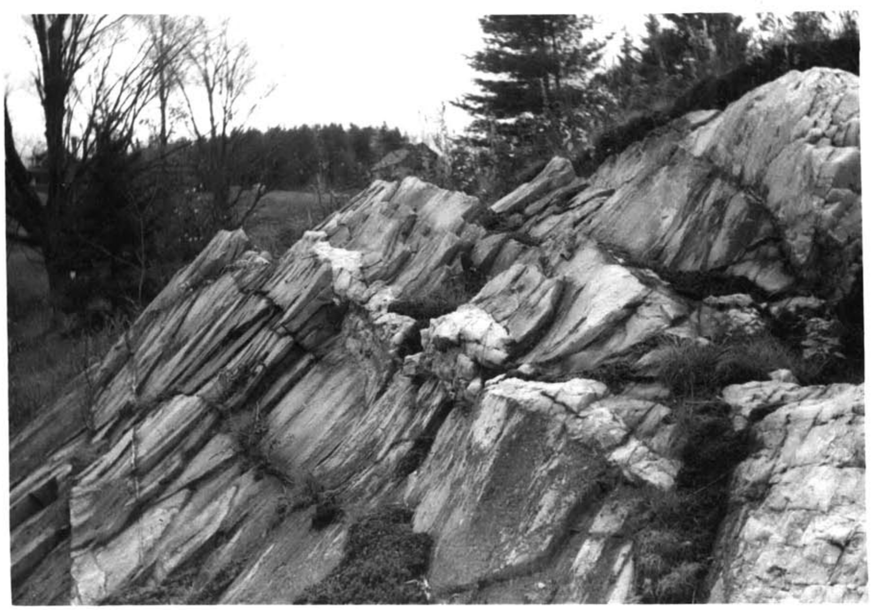
The locality is on the northeast side of Vermont Highway 106, southwest of Wardner Hill in Reading. The boudinage occurs in a sill-like mass of granitic rock enclosed in banded biotite-muscovite gneiss. The filling is of quartz. (Photograph by H. S. Yoder)

(B)

Quartz-filled boudinage in quartzites of Plymouth member of Grahamville formation, near the fold in Plate IX-B. The boudinage is parallel to the fold axis.



A



B

Mountain massif. The interbanded schists and amphibolites of the Pinney Hollow formation in the Chester dome probably afford the most striking examples of boudinage in various stages of formation. Different occurrences of boudinage, however, vary considerably as to the shape of the cavity resulting from the rupture of the competent bed. The concavity of the ruptured edges in amphibolite boudinage indicates considerable plasticity even in the relatively competent amphibolite. Boudinage of more brittle materials such as quartz veins and pegmatite dikes generally does not show this feature. The cavities formed by boudinage are filled partly by flowage of the more plastic material and partly by quartz and other secondary minerals.

The orientation*of the boudinage in the Chester dome is generally radial. The exceptions to this rule are those boudins occurring on the short limbs of the reversed drag-folds. Boudinage of the latter type is generally oriented parallel to the fold axis. In the younger rocks on the eastern flank of the Green Mountain massif, the boudinage is normal to the fold axes except in the region west of the Plymouth-Ludlow Valley where the few observed occurrences are parallel to the fold axes.

Relative plasticities of rock-types

Boudinage affords an excellent indication of the relative plasticities of the various rocks involved at the time of deformation. The following list gives the more

*Next page.

common rock types in the order of increasing plasticity as determined from the observation of many different occurrences of boudinage:

1. Vein quartz and pegmatite
2. Quartzite
3. Hornblendic gneisses
4. Micaceous gneisses
5. Hornblende schists
6. Mica schists
7. Carbonate rocks

The relations between the various schists and gneisses are reversed in some instances. Most of such reversals appear to be related to pronounced differences in grain size, with the finer-grained rock the more plastic. The carbonate rocks, however, appear always to be more plastic than any of the other types with which they have been seen in contact, regardless of grain size.

The relative plasticities determined from observation of the deformed pebbles in the Tyson conglomerate are in essential agreement with those based upon boudinage except that fewer rock types could be compared. The squeezing of the schistose beds of the Reading gneiss into the crests and troughs of the minor folds is a further indication of the relative plasticity of schists and gneisses.

*By "orientation" or "direction" of boudinage the writer means the direction of the lines of rupture separating the boudins.

The correlation in the silicate rocks, of coarseness of grain with plasticity is significant and suggests that the deformation in such rocks is accomplished largely by intergranular rather than by intra-granular movement, inasmuch as fine-grained rocks have a relatively greater "acreage" of inter-granular surface per unit volume. The extreme plasticity of the carbonate rocks, however, relative to the silicates indicates that intra-granular deformation is probably of much greater importance in carbonates. The greater plasticity of micaceous rocks as compared with non-micaceous rocks of comparable grain size might suggest that intra-granular deformation is also important in rocks rich in the micaceous minerals. It is equally probable, however, that the relative ease of inter-granular movement in rocks composed largely of flat platy minerals is sufficient to explain their greater plasticity.

Analysis of the Deformation

Theoretical considerations

The relations between the various minor structural features and the overall deformation of the rocks provide many interesting problems. The aim of most minor structural studies, including this one, is to obtain evidence concerning the deformation in a given area beyond that which is evident from the analysis of the large scale structures. Rosenfeld (personal communication) has suggested that structural features that are compatible with one deformation, and hence can be related to one strain ellipsoid be called "congruent" structures. That two structural features can be related to the same strain ellipsoid does not necessarily mean that they were actually produced by the same deformation, but does show that they could have been. If a rock shows a large number of such features all of which are mutually congruent, however, it is highly probable that the structures are the result of but one deformational process. All of the large scale features* in the area mapped by the writer appear to be compatible with one deformational process, but the relationships of most of the small scale features are less obvious.

The strain ellipsoid, as applied to the study of a permanent, non-elastic deformation, may be defined as the

*Except, of course, those in the basement complex.

final shape assumed by an initially spherical region within the deformed mass. Such a shape will be a true ellipsoid, however, only if the deformation is homogeneous. It is commonly assumed that the deformation of rock masses is homogeneous, but it is clear that this assumption is not valid in the general case and that conclusions based upon it must be accepted with some reservations.

An unfortunate feature of the strain ellipsoid is that although it may describe correctly the end product of any homogeneous deformation, it gives almost no information concerning the actual movements by which the deformation was produced. In the present area, however, the rotation of porphyroblasts and other features indicate that the foliation planes were definitely surfaces of differential movement during the deformation. It might therefore be assumed, as a first approximation, that as much of the deformation as is geometrically possible actually did take place by slip along the foliation planes, the remainder to be accounted for by movements of another sort.

The only possible components of a homogeneous deformation that cannot be accounted for by pure laminar slip upon one set of parallel planes are changes in shape of an initially circular area in the plane of slip and elongation or shortening normal to the plane of slip. Any deformation by pure laminar slip may be described by an ellipsoid with one of its circular sections parallel to the slip planes and in which

B is equal to the radius of a sphere of the same volume as the ellipsoid ($B = \sqrt{AC}$). B must coincide with b and A must be oriented so that a rotation of forty-five degrees or less in the sense of the movement will bring it into coincidence with a. The components of the deformation that cannot be accounted for by laminar slip may be described by an ellipsoid with one of its principal axes normal to the slip planes and the other two axes parallel to the slip planes. There is no necessary specification as to which axis is normal to the slip planes, and no necessary relationship between the axes paralleling the slip planes and the movement axes a and b. Such an ellipsoid may be regarded as a "partial" ellipsoid representing the shape assumed by an initially spherical region in the deformed mass as the result of a partial deformation, and the actual strain ellipsoid as the sum of the partial operations acting either simultaneously or in sequence upon the same initially spherical region.

Even though the assumptions concerning the actual movements cannot be fully justified, it is correct to state that the partial ellipsoid describing all components of the deformation that could possibly have resulted from laminar slip on the foliation planes represents the tangential strain relative to the foliation planes, and that the partial ellipsoid describing all other components of the deformation represents the normal strain relative to the foliation planes. Inasmuch as the A and C axes of the ellipsoid of tangential strain must rotate as deformation proceeds, relative to the plane of foliation, this component of the deformation will be referred to henceforth as the rotational component. The ellipsoid of normal strain, on the other hand, must always have one axis normal to the foliation plane, and will be referred to as the non-rotational component, despite the fact that rotation of the axes in the foliation plane about the normal axis may occur in some instances.

A complication in the analysis of rock deformation is the non-homogeneity of the material being deformed. Fortunately, however, the problem is somewhat simplified by the fact that the inhomogeneities are commonly of a laminar nature and the laminae, at least in the present area, generally parallel to the foliation. It is obvious that adjacent beds or bands, differing greatly in competency, will not respond equally to a given deforming force, and that a slightly different strain ellipsoid will be required to describe the deformation in each type of bed, with that in the more competent beds generally the more nearly spherical. It is instructive to consider such differential deformation further with respect to the rotational and non-rotational components. As far as pure laminar slip parallel to the banding is concerned, there seems to be no great difficulty involved in strong differential effects inasmuch as the areas of bands or beds in contact with each other remain constant. In an extreme case the deformation might be regarded as restricted entirely to the relatively incompetent layers, the competent layers behaving like rigid plates separated by the incompetent lubricant. In elongation or shortening normal to the banding, however, difficulties arise in that differential thickening or thinning of adjacent laminae also involves changes in their relative areal extent. If the laminae are thickened, the incompetent layers should show a decrease in area relative to the competent layers in the event of differential effects between the two layers, and the reverse should be true in the case of thinning. In the first instance geometric compensation might be attained by buckling or imbrication of the competent layer, and in the second instance by boudinage of the competent layer. Another possible type of compensation exists in closely folded rocks where differential thickening and thinning in different parts of folds may be, in part, mutually compensated by flow of the incompetent material away from the limbs and toward the crests and troughs. It is significant, however, that all of the various methods of compensation would tend to be resisted by various inherent physical properties of the rocks such as the shearing and tensile strength of the competent layer, or, in the case of flowage in folds, the viscosity of the incompetent layer. This should mean that differential effects between layers of varying competency should tend to be more marked in the case of pure laminar slip parallel to the banding than in elongation or shortening parallel to the banding because of the lag in the necessary compensations. The same arguments should, of course, apply to the rotational and non-rotational components of a more generalized deformation insofar as the rotational component actually does represent slip parallel to the banding and foliation. Although the deformation might conceivably be entirely eliminated from the competent layers as far as the rotational component is concerned, the writer doubts that this could ever be true with respect to the non-rotational component.

Rotational features

The important rotational minor structures, described in earlier sections, include small asymmetric folds, "snowballed" porphyroblasts, and slip-cleavage. The small asymmetric folds and the rotated porphyroblasts appear, within the limits of observation, to be consistent with the larger scale rotational features and the strain ellipsoids deduced from them. About the only questionable feature is the extreme contortion of the short limbs of many of the drag folds in the Chester dome area. The writer interprets this, however, as indicating that in its rotation from the plane of the long limb to its present position the short limb has suffered a reversal of shear sense, and that this process of "rubbing the fur the wrong way" has produced the extreme contortion.

As has been mentioned in a preceding section, asymmetric crinkles, the rotation of porphyroblasts and other features indicate that the foliation has served as a plane of slip during the deformation. Slip-cleavage complicates the picture in that its formation introduces a second set of slip-planes, upon which the movements are in the same rotation sense and about the same b-axis as those upon the foliation planes from which the slip-cleavage is derived. Turner (1948, pp. 169-170) interprets such movements as incompatible with a single deformation, apparently on the assumption that if simultaneous or nearly simultaneous movements occur on two sets of intersecting slip-planes, the rotation sense of the two movements must be opposite. According to the writer's analysis, however, (Figure 6) Turner's interpretation should apply only to slip-planes lying in different quadrants of the ellipsoidal section normal to b, although movements in either sense might be permissible for planes parallel to one of the principal axes of the ellipsoidal section. That the development of slip-cleavage is incompatible with the movements upon the foliation planes seems erroneous to the writer inasmuch as these movements are directly responsible for the formation of the slip-cleavage in the first place. The mechanism appears to be a small-scale replica of that involved in thrust faulting.

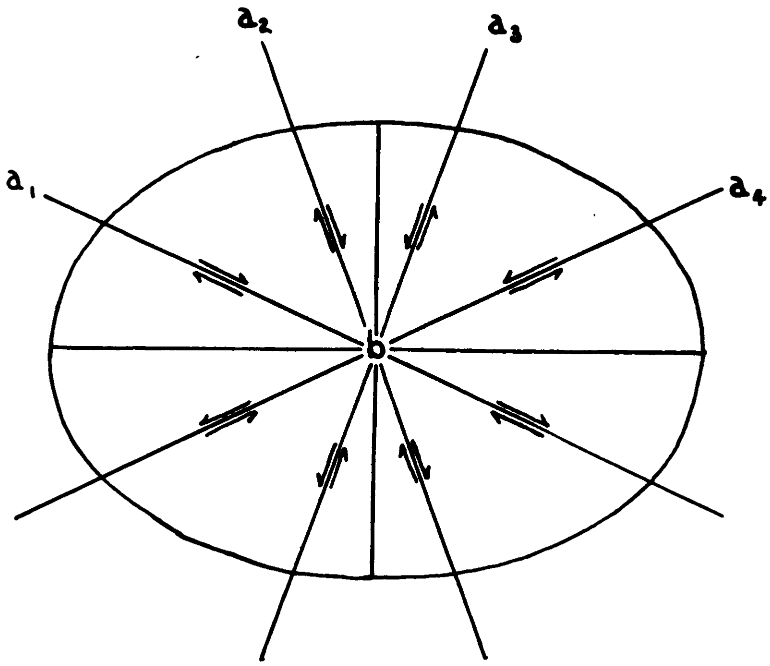


Figure 6. Rotational movements and the strain ellipsoid.

The ellipse represents the section of the strain ellipsoid normal to b. Simultaneous infinitesimal movements of rotation-sense other than those indicated by the arrows, for various positions of a, would reduce the eccentricity of the ellipsoidal section and be incompatible with the ellipsoid.

The fact that the slip-cleavage on the opposite limbs of small folds is generally parallel, but with opposite shear sense, indicates (Figure 6) that it must coincide with one of the principal sections of the ellipsoid, presumably the AB plane, and that it is probably safe to regard most slip-cleavage as essentially an axial plane phenomenon in the interpretation of field data.

Tensional features

The tensional features studied by the writer include boudinage and joints. Of the two, boudinage lends itself most readily to analysis by virtue of its being essentially a two-dimensional phenomenon. It is noteworthy, however, that the two most prominent joint systems (normal to the fold axes, and at right angles to this, roughly normal to the dip) include the two chief directions* of boudinage (parallel and perpendicular to the fold axes) indicating that the two tensional phenomena are probably compatible with each other.

The significance of boudinage is twofold: It clearly indicates that the deformation has involved flattening of the rock layers, the boudinage representing an "attempt" of the competent bed to increase its area in accord with that of the incompetent bed; and, by its direction,* it shows the directions of relative elongation and shortening in the bedding planes. These directions of elongation and shorten-

*See footnote Page 109

ing should be the major and minor axes, respectively, of the section of the strain ellipsoid parallel to the bedding planes. In Figure 7, B is assumed parallel to the fold axes. The bedding planes on different parts of the folds may show all possible orientations passing through B. The position of the major axis of the ellipsoidal section parallel to the bedding will then be related to the orientation of the bedding planes relative to the circular sections (BB') of the ellipsoid. Inasmuch as the bedding planes always include B, it must always be equal to one axis of the ellipsoidal section. For planes between AB and the circular sections B will clearly be the minor axis of the ellipsoidal section and boudinage should develop parallel to it, and for planes between AC and the circular sections the reverse should be true. It is apparent from this analysis that the occurrence of boudinage in the Chester dome area is compatible with an ellipsoid oriented as in Figure 8. The same relationship also provides convincing evidence that B is, indeed, parallel to the fold axes in this specific instance, because if A were the fold axis, all boudinage would have to be normal to it, and if C were the fold axis, all boudinage would have to be parallel to it.

Boudinage is also significant in that it is strictly a function of those components of the deformation that <u>cannot</u> be related to laminar slip parallel to the bedding,

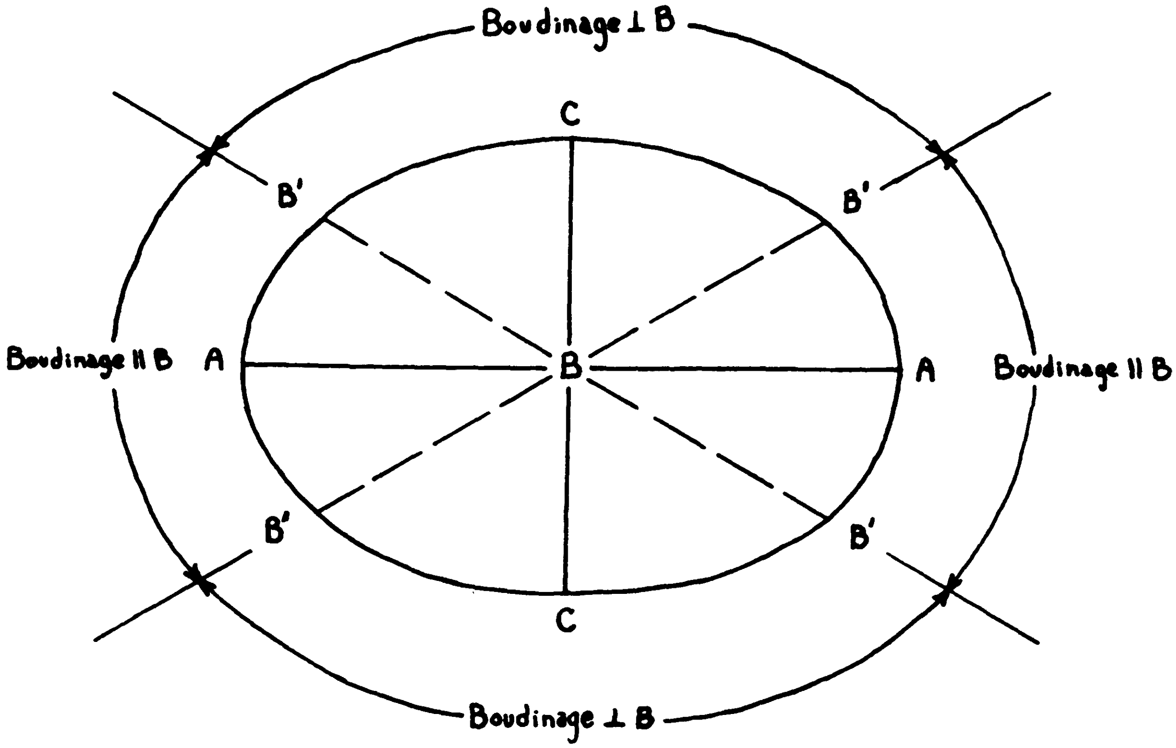


Figure 7. Boudinage orientation and the strain ellipsoid.

If the fold axes are parallel to B, the boudinage orientation is determined by the position of the bedding planes relative to the circular sections, BB'. The boudinage must be normal to the long axis of the ellipsoidal section parallel to the bedding planes.

specifically, the thinning of the beds, and the elongation and shortening in the plane of bedding. It thus gives us our first specific information on the orientation of the non-rotational partial ellipsoid. It is evident that the normal to the bedding must be an axis of shortening for boudinage to exist and therefore either C or B, most probably C of the partial ellipsoid, and that the normal to the boudinage in the bedding plane must be A of the partial ellipsoid. It is interesting to note that A of the non-rotational partial ellipsoid is parallel to B of the "total" ellipsoid under the conditions whereby boudinage occurs perpendicular to the fold axes.

Compressional features

Compressional features are, of course, corollary to tensional features and the relations that must hold between them are clear-cut. Simple symmetrical folds are probably best classified as compressional features but are not characteristic of the area under discussion. The non-rotational folds plunging about down-dip just north of the Dry Hill gneiss-block in Plymouth, however, are possibly compressional in origin, an interpretation compatible with the fact that the few occurrences of boudinage west of the Plymouth-Ludlow valley are approximately normal to them and parallel to the axes of the clearly rotational folds in the same area. Inasmuch as the axes of these non-rotational folds are approximately parallel to the inferred direction of movement, they are "a" folds as described by Cloos (1946, p. 26). Such folds are not uncommon in the vicinity of large thrust faults and may possibly be related to convergent movements along the fault plane. The Dry Hill folds are quite definitely in an area of thrusting and are similar in appearance to the "a" folds in the Canajoharie shales beneath the sole of the Champlain thrust in northwestern Vermont.

Pebble deformation

The deformed pebbles of the area mapped are interesting in that their axes do not conform, in any instance yet observed, to the strain ellipsoid axes deduced from other criteria. West of the Plymouth-Ludlow valley slip-cleavage and other features indicate that the AB plane of the strain ellipsoid is about vertical with B parallel to the gently north-plunging fold axes. The intermediate axes of the pebbles are parallel to B but the long axes are in the foliation plane and, therefore, generally at about 45 degrees to A. In the Chester dome area, and in the Shaw Mountain conglomerates on the western limb of the Proctorsville syncline, the strain ellipsoid is oriented approximately as in Figure 10, but the pebbles are elongate parallel to the fold axes (B of the strain ellipsoid) and flattened parallel to the bedding foliation.

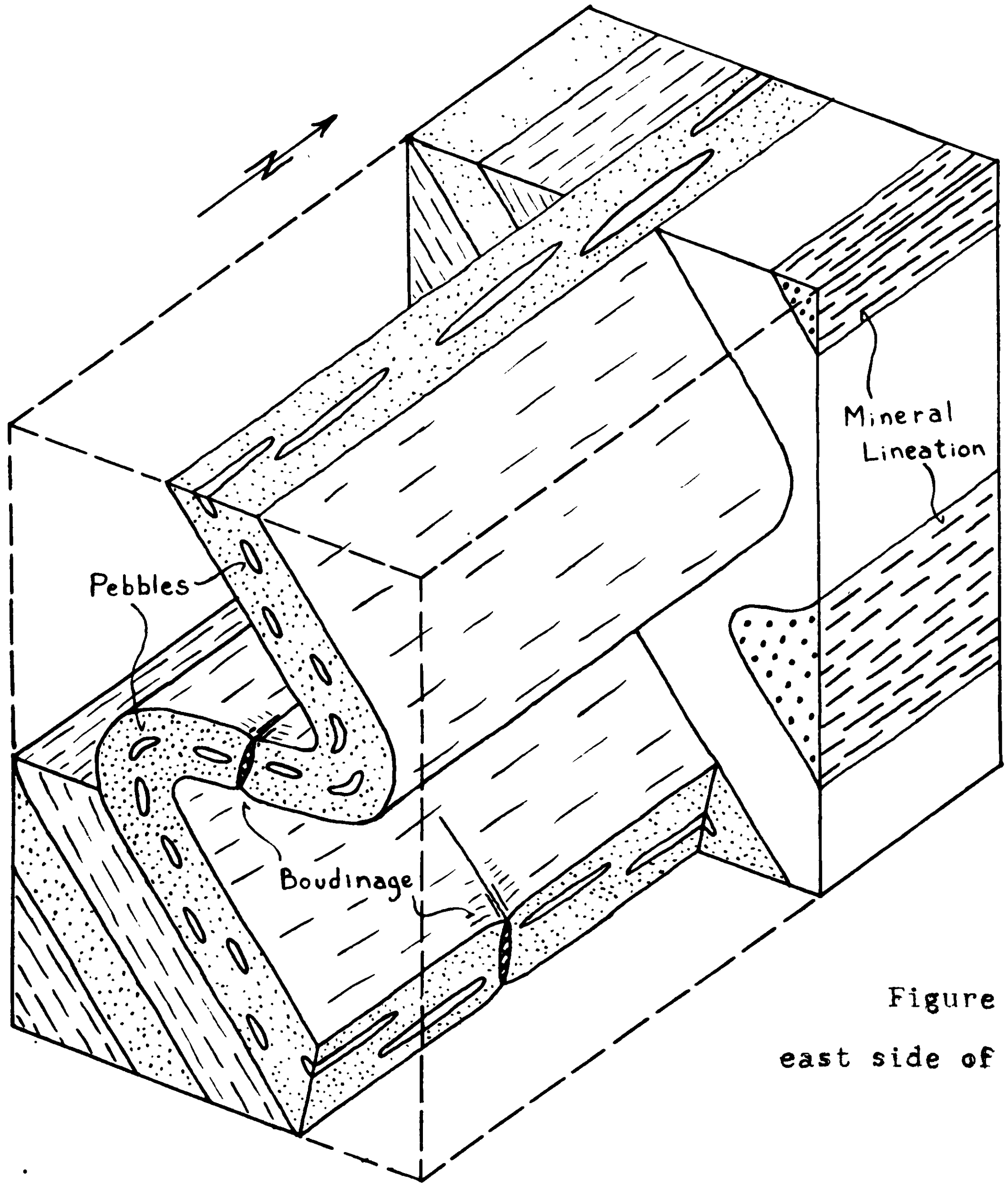
It is, of course, possible that the deformed pebbles are incompatible with the rest of the deformation, but in the apparent absence of other incompatible features the writer does not consider this likely, particularly in view of the fact that pebble deformation in other localities shows similar peculiarities. The writer suggests, as an alternative, that many of the apparent anomalies are related to the fact that the conglomerate horizons are among the most highly competent beds and that their deformation should, therefore, not be typical of that of the rock mass as a whole. As has been pointed out in an earlier section, the deformation of the relatively competent layers should differ from that of the rest of the rock mass in being less intense and in that the laminar-slip component should be relatively less important. This should mean, furthermore, that the strain axes in the competent bed should tend to be similarly oriented to those of the non-rotational partial ellipsoidal. Assuming now, merely that the pebble axes will be oriented similarly to the strain ellipsoid for the bed in which they occur, it is possible to test the hypothesis, using the position of boudinage to orient the non-rotational partial ellipsoid. In the Chester dome the boudinage, except on the short limbs of drag folds, is normal to the fold axes, indicating that except for the short limbs of folds, the non-rotational partial ellipsoid is oriented with its long axis parallel to the fold axes and short axis probably normal to the bedding, and therefore in agreement with the pebble axes. At the Springfield locality the conglomerate has been folded, so that some of the pebbles are now on the short limbs of the folds with the "wrong" orientation for that limb, but the fact that the pebbles are themselves folded indicates that their flattening almost certainly took place before they became involved in the drag fold. West of the Plymouth-Ludlow valley the hypothesis also checks inasmuch as the prevailing boudinage is parallel to the fold axes, and normal to the pebbles which are flattened in the plane of bedding. The deformation of pebbles, therefore, seems to support the assumption that deformation by laminar slip tends to be eliminated in the relatively competent beds.

Mineral orientation

The significance of the parallel orientation of minerals has long been a moot point among structural geologists, particularly with respect to foliation. Fairbairn (1935) has outlined the problem and has shown that different mineral orientations with respect to the strain ellipsoid can occur and that the different types of orientation are largely functions of the nature of the deformation. It seems clear that in pure laminar slip randomly oriented platy minerals would eventually be rotated into the plane of slip, which should be the only stable position. It also seems clear that in a non-rotational deformation, with no dominant planes of slip, randomly oriented platy minerals should be rotated as deformation proceeds into the A-B plane of the ellipsoid, the "plaiting" surface of Fairbairn. In the first case, all plates would rotate in the same sense, that of the laminar slip, and in the second case most of the plates would rotate about different axes and in different senses, depending upon their initial orientation. It seems to follow, therefore, that orientation of platy minerals relative to the strain ellipsoid is primarily dependent upon the actual movements involved in the deformation. In terms of the deformation components proposed in the preceding section, it seems reasonable to suggest that if the rotational (laminar-slip) component is dominant, the foliation should develop parallel to *a*-*b*, the position of which should be related to that of initial planes of weakness in the rock, presumably parallel to the stratification, and that if the non-rotational component is dominant the foliation should develop parallel to A-B, and, in the general case, should be inclined to the stratification planes.

In the present map-area the prevalence of bedding foliation would thus indicate that a major or at least considerable part of the deformation can be related to laminar slip parallel to the stratification planes. That the foliation planes are primarily planes of movement, and not plaiting surfaces, is supported by the presence of "snowballed" porphyroblasts and numerous small asymmetric crinkles.

The orientation of rod-shaped minerals is analogous to that of platy minerals in most respects. Deformation by laminar slip should rotate randomly oriented rods toward the plane of slip and also toward *a*. The tendency to approach *a*, however, would decrease to the extent that the rods were initially sub-parallel to *ab*, and would be nil for rods already in *ab*. This would result in an imperfect mineral lineation parallel to *a*. The writer interprets the streaming on the eastern limb of



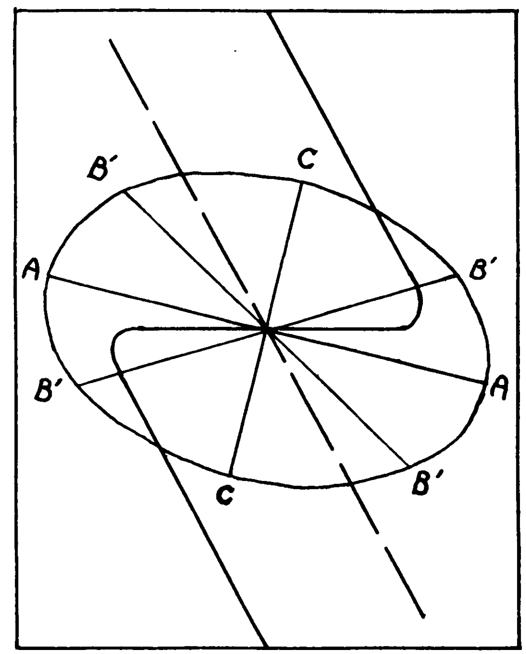


Figure 8. Minor structural relations, east side of Chester dome.

the Green Mountain anticlinorium as of this type, a conclusion supported by the fact that it is always normal or sub-normal to observed rotational axes.

The orientation of rod-shaped minerals by non-rotational deformation should be parallel to A. The only conspicuous mineral lineation in the Chester dome is that of the hornblende in the amphibolites and hornblende gneisses. The ambibolites are generally more competent than the rocks with which they are interbedded and show some of the best examples of boudinage in the area. As might be predicted from the orientation of deformed pebbles in the same region, the mineral lineation is normally perpendicular to the boudinage in the same bed and, therefore, parallel to A of the non-rotational partial ellipsoid. Exceptions have been noted on the short limbs of some drag folds, again probably indicating relatively late formation of the drag. On at least one short limb hornblende crystals have been observed to run both ways. The conditions controlling both mineral and pebble elongation in relatively competent beds appear to be the same and related to the non-rotational components of the deformation.

Summary

By way of summary it may be stated that none of the small-scale structural features observed by the writer in the younger rocks appears to be incompatible, assuming that the analysis is valid, with the rest of the deformation. This

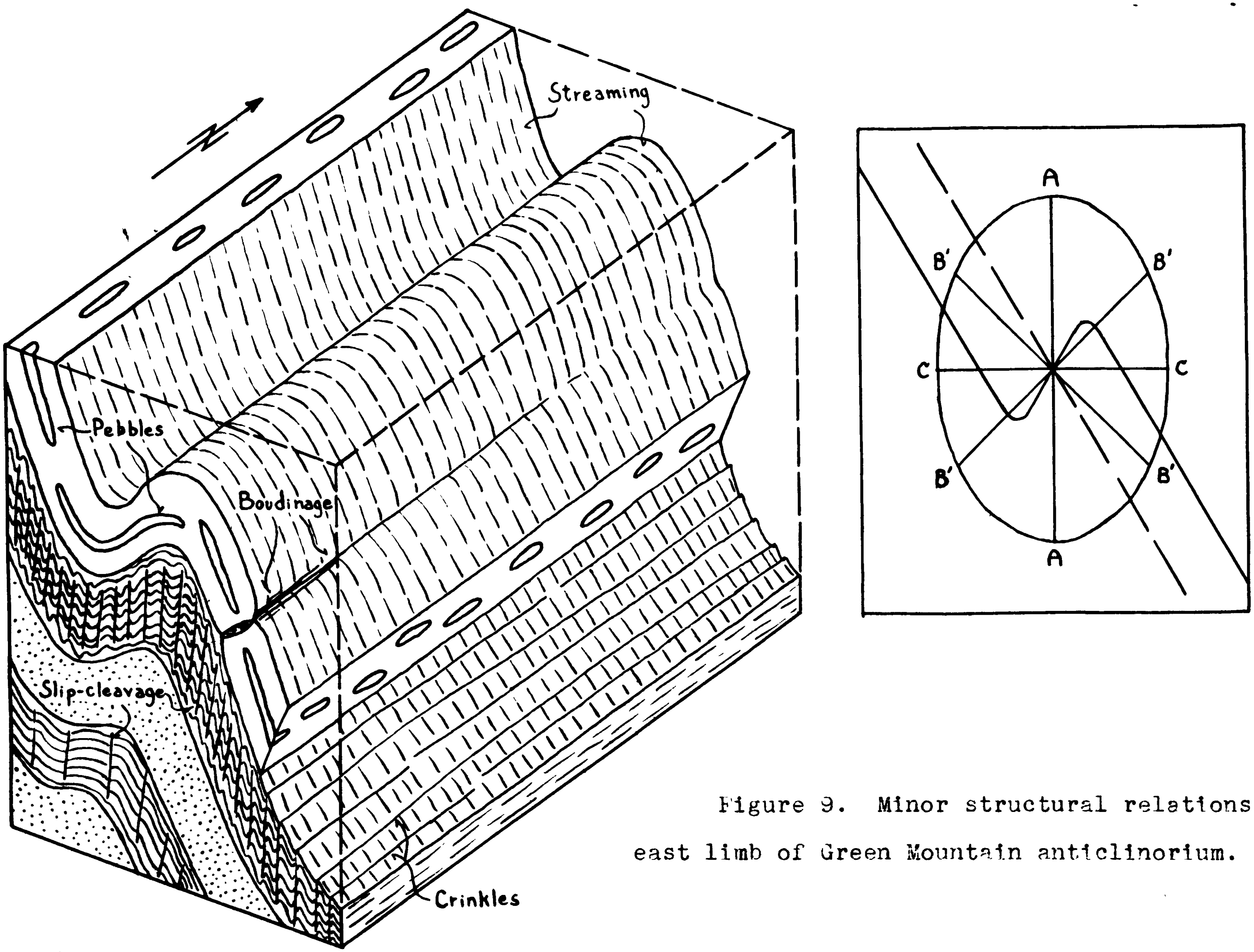


Figure 9. Minor structural relations, east limb of Green Mountain anticlinorium.

does not, of course, constitute hard and fast proof that only one deformation has taken place, but does indicate, if more than one has taken place, that the large-scale movements were of the same general nature, so that the effects of the different deformations are indistinguishable. Features such as the offset of the axial plane of the Proctorsville syncline, the later folding of flattened pebbles and the offsetting of foliation by later slip cleavage are, to be sure, records of successive deformational events, but the writer can see no valid reason why they are not to be regarded as stages in one continuous process, related to the same movements and related to the same causal forces.

The relationships between most of the minor structural features in the Chester dome area have been indicated in Figure 8. In Figure 9 the corresponding relationships are shown for the eastern limb of the Green Mountain anticlinorium in the region just west of the Plymouth-Ludlow valley. East of the valley the slip-cleavage acquires an increasingly gentle westward dip and boudinage is perpendicular rather than parallel to the fold axes.

In Figure 10, the approximate orientation of the strain ellipsoid at different points in the Green Mountain and Chester dome area is shown as inferred from the various structural data. The apparent transition between the Chester dome and Green Mountain deformation is interesting in view of the fundamental differences between the two structures, the

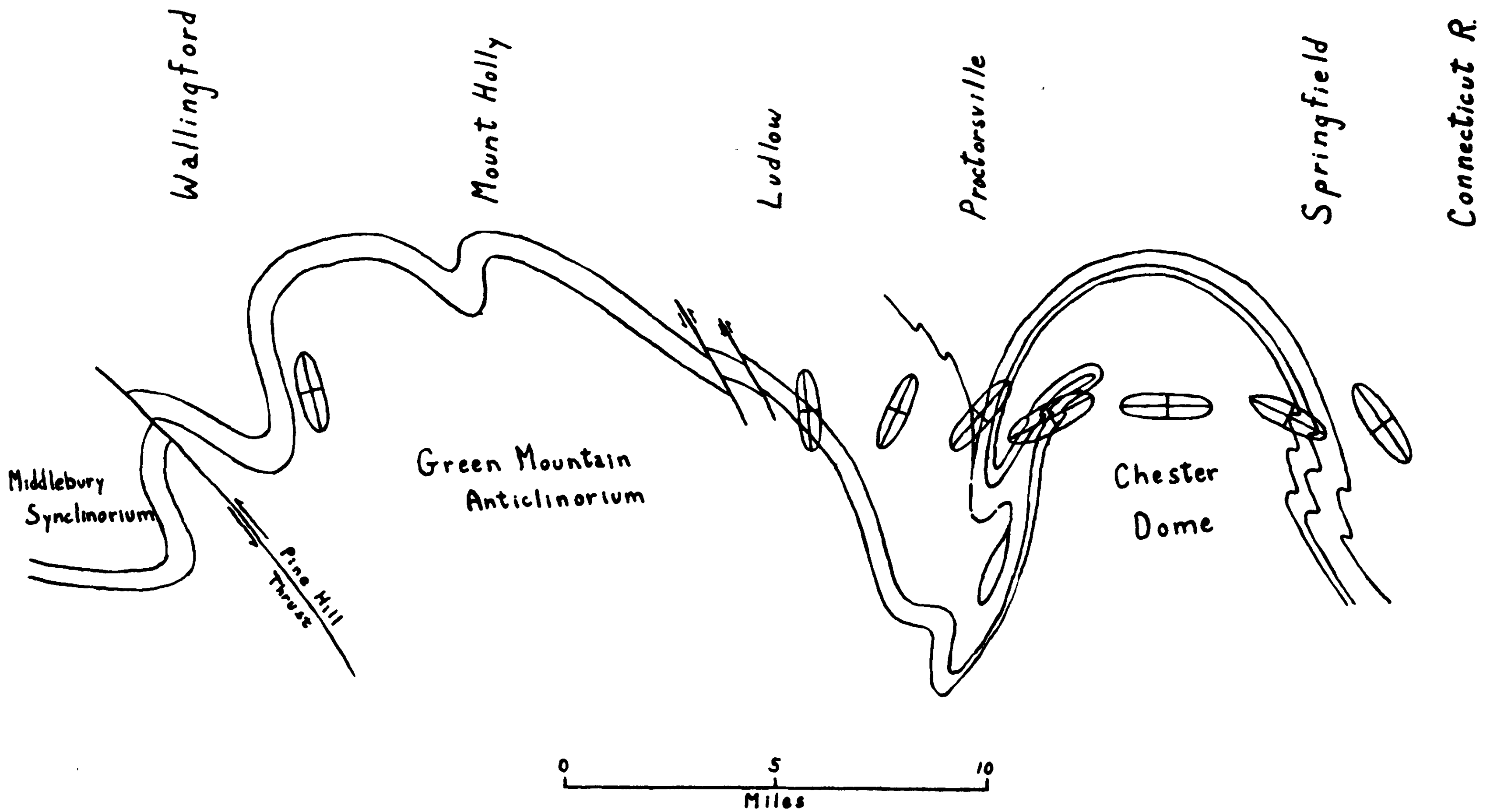


Figure 10. Orientation of strain ellipsoid, Green Mountain anticlinorium and Chester dome.

Green Mountain anticlinorium apparently related to lateral, and the Chester dome to vertical movements.

METAMORPHISM

Regional Metamorphism

Metamorphic zoning

The isograd map (Figure 11) indicates that there is a fairly distinct metamorphic zoning in the vicinity of the Chester dome. The isograds are based primarily upon the appearance of key minerals in schists of argillaceous composition, and divide the area into three zones which have been called, proceeding from the areas of relatively low grade metamorphism toward the center of the dome; the chlorite zone, the garnet zone and the staurolite-kyanite zone. The staurolite-kyanite zone is clearly limited to the immediate vicinity of the dome but the limits of the garnet zone to the north and south are not known. In addition to the increase in metamorphic grade toward the dome there is also a progressive increase in grain size, the rocks of the chlorite zone being fine-grained and phyllitic, and the schists of the staurolite-kyanite zone containing staurolite and kyanite blades up to two inches in length and garnets up to an inch in diameter.

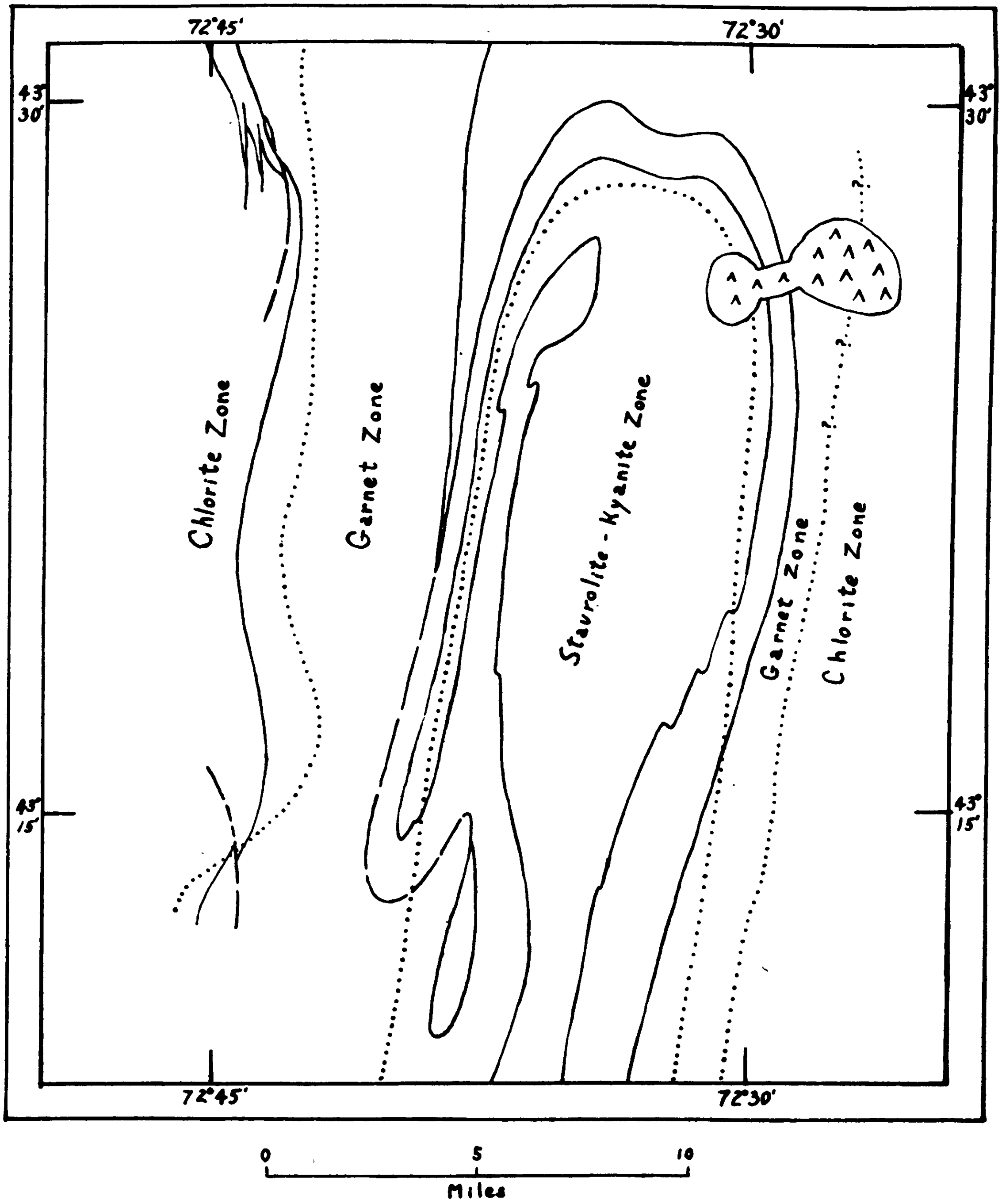


Figure 11. Metamorphic zoning.

Dotted lines show approximate locations of main isograds. (Biotite zone not differentiated.)

The writer has been unable to differentiate a biotite zone, as is commonly done, and believes the distribution of biotite to be more a function of composition than grade of metamorphism. The presence of biotite appears to be related to a low alkali-alumina ratio, inasmuch as muscovite and chlorite appear to form in its stead in rocks with excess alumina (Figure 12). In rocks with a high alkali-alumina ratio, such as the arkoses and greywackes of the Tyson formation and Mendon series, biotite is persistent through all grades of metamorphism in the area mapped by the writer, appearing in abundance even on the western limb of the Green Mountain anticlinorium in certain of the arkoses and greywackes of the Mendon series. It is also persistent in impure dolomitic rocks, minute scales of phlogopite appearing even in the Rutland dolomite. In certain rocks of more argillaceous composition and, therefore, a lower alkali-alumina ratio, however, it is true that there is a sudden "blossoming" of biotite corresponding rather closely with the first appearance of garnet. Such "late" biotites are commonly iron-rich, occurring as conspicuous rhombic plates randomly oriented, in many instances, relative to the schistosity. The writer suspects that this more conspicuous occurrence of biotite probably represents the biotite zone of Harker (1932).

There are several possible explanations for this delayed appearance of biotite. It is clear from Figure 12 that the average argillaceous rock has a composition close to the muscovite-garnet (or chlorite) join and that a slight unbalancing

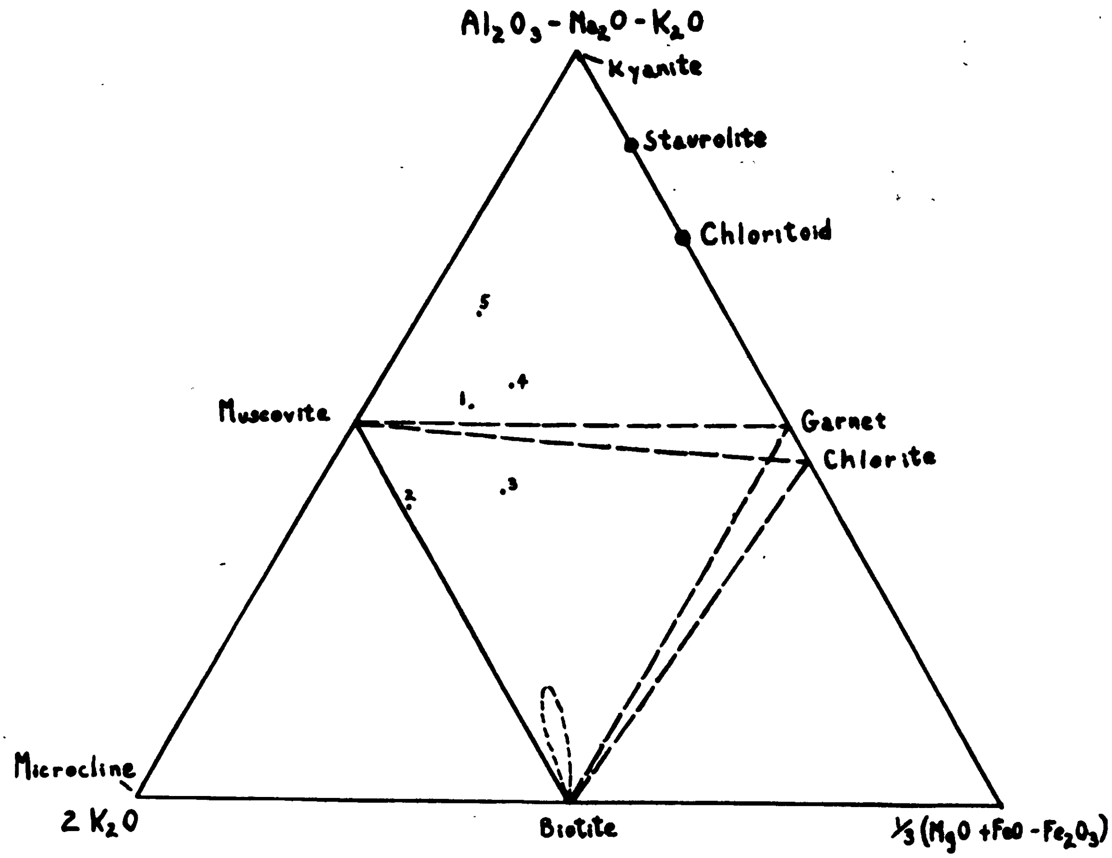


Figure 12. Composition diagram for argillaceous and arenaceous sediments.

The diagram is plotted upon the basis of the moleeular proportions of the major constituents. The corners of the triangle are so weighted that the position of a rock in the muscovite-biotite-garnet (or chlorite) field shows the approximate relative proportions of the minerals by volume. Ferric iron is assumed present as magnetite, and soda as albite. In highly aluminous types, however, soda is actually present as paragonite.

1. Average shale (Clarke, 1924, p 34)
2. Green slate, Pawlet, Vt. (Clarke, p 554)
3. Purple slate, Castleton, Vt. " "
4. Schist with 11% albite (Barth, 1936)
5. Middle Cambrian shale, Alabama (Clarke, p 552)

of the mineralogic relationships might well throw such a rock into the biotite field. One obvious explanation is that rocks showing late biotite may have compositions lying in the wedge-shaped area bounded by muscovite, chlorite and garnet, so that formation of the more aluminous garnet at the expense of chlorite would require the simultaneous appearance of biotite. Although this explanation is qualitatively valid and fits most of the facts it seems rather remarkable that so many argillaceous rocks should have had such a restricted composition. Other plausible explanations include metasomatic addition of alkali, magnesia or ferrous iron, possibly combined with loss of alumina, but in the absence of other evidence of large-scale metasomatism, the writer doubts the validity of such explanations although they are almost impossible to disprove. Still another plausible explanation, however, is the reduction of ferric iron to ferrous with rising metamorphic grade, possibly as a function of the presence of graphite. This fits most of the facts outlined above, as well as the apparent mutual exclusiveness of graphite and ferric iron-minerals in the higher metamorphic grades, and the fact that such late biotite has its most outstanding development in the Whetstone Hill member of the Moretown formation, a horizon notable for its extremely erratic graphite-distribution, both along and across strike, as well as for its highly ferruginous nature. This last hypothesis, also supported by the iron-rich nature of the late biotites, is the one favored by the writer, coupled with the effects of the chlorite-garnet reaction. The erratic distribution of graphite in the Whetstone Hill member of the Moretown formation would thus be interpreted as the result of the removal of the carbon, probably as CO_2, with attendant reduction of ferric iron to the ferrous state.

The writer also deviates from most classical treatments in not mapping a separate kyanite zone representing a higher grade of metamorphism than the staurolite zone. The presence of kyanite, however, requires an unusually high alumina content, such as might be expected only in a relatively pure clay shale, suggesting that the non-appearance of kyanite in many staurolite schists may be of doubtful significance. The fact that staurolite contains crystal-structural units which are, to all intents and purposes, kyanite, seems incompatible with any significant difference in stability range as a function of pressure and temperature. Their occurrence in parallel growths upon their common structural plane, (100) of kyanite and (010) of staurolite, moreover, seems inconsistent with anything but contemporaneity of formation.

The fact that garnet seems to form at the expense of chlorite does not, as is commonly assumed, indicate almost complete compositional equivalence of the two minerals. Chlorite persists, apparently as a stable constituent, well into the staurolite-kyanite zone, occurring as fresh porphyroblasts, clearly unrelated to equally fresh garnet and biotite occurring in the same rock. Chlorite rims around garnet, or pseudomorphs after garnet, have been observed, particularly in the Green Mountain basement rocks, and are probably indeed retrograde, but should not be taken as evidence that other associations of chlorite and garnet (or chlorite and biotite) have the same significance. To be specific, the garnets of regionally metamorphosed argillaceous rocks (Weiss, 1949) are notably non-magnesian, whereas the chlorites (aluminous prochlorites) are distinctly so, with iron-magnesium ratios rarely in excess of 1 : 1. The chlorites in the present area show a progressive drop in index of refraction with increasing metamorphic grade, whereas the garnets stay about the same, indicating that they are probably formed at the expense of the iron-component of the chlorite rather than the chlorite as a whole.

Comparison of the greenstones and amphibolites with the adjacent schists indicates that the chlorite zone corresponds essentially to the greenstone-schist facies of Eskola (1920), and the garnet and staurolite-kyanite zones to the epidote-amphibolite facies. In the depth-zone terminology of Grubenmann and Niggli (1924), the chlorite zone would be the "epizone" and the higher grade zones, parts of the "mesozone". The writer believes, however, that the relations between depth, temperature and pressure, implied by the Grubenmann-Niggli system, are not justifiable.

Metamorphism of arenaceous-argillaceous sediments

Typical mineral assemblages in metamorphosed arenaceous-argillaceous rocks of the chlorite zone are, with quartz and minor constituents omitted, and calcium-rich types not considered for simplicity:

1. Microcline-albite-biotite.
2. Microcline-albite-biotite-muscovite.
3. Albite-biotite-muscovite.
4. Albite-biotite-muscovite-chlorite.
5. Albite-muscovite-chlorite.
6. Muscovite-chlorite.
7. Muscovite-chlorite-chloritoid.

The listing is in approximate order of decreasing alkali-alumina ratio with compositional types corresponding to arkoses and greywackes toward the top, and shales toward the bottom. In the garnet zone, the first three assemblages are unchanged, garnet appearing as an additional constituent in the last four. Chlorite tends to persist, particularly in the aluminous types, but may be absent in some instances from assemblage (4). . In the lower staurolite-kyanite zone the first six assemblages are as in the garnet zone but the more aluminous types must be revised as follows with the addition of an extra assemblage:

7. Muscovite-chlorite-garnet-staurolite.
8. Muscovite-chlorite-garnet-staurolite-kyanite.

Higher in the staurolite-kyanite zone the stability range of even magnesian chlorite is apparently exceeded and the more aluminous assemblages must be revised radically as follows:

4. Albite-biotite-muscovite-garnet.
5. Biotite-muscovite-garnet.
6. Biotite (phlogopite)-muscovite-garnet-staurolite.
7. Biotite (phlogopite)-muscovite-garnet-staurolite-kyanite.

It is clear that the relatively aluminous rocks are the more grade-sensitive. The absence of albite in the more aluminous types is interesting in view of the fact that argillaceous rocks without a significant soda-content are almost unknown. Partial analyses, however, of the mica from the Gassetts schist (Currier, 1934) show that the soda is present as a paragonitic component in the white mica. It is reasonable to infer, therefore, that in other albite-free mineral assemblages the soda is probably also present as a paragonitic component of the muscovite. The association of albite with staurolite and even kyanite has been reported by Rosenfeld (personal communication), indicating that the stability range of paragonite must be exceeded in the staurolite-kyanite zone under some conditions. Such mineral assemblages, however, have not been observed in the areas mapped by the writer. It is unfortunate that the soda-content does not affect the optical properties of muscovites significantly. Another, probably related, problem is the disposition of the excess alumina in the lower grade equivalents of kyanite-bearing schists. It is reasonable to conjecture that it may be present as pyrophyllite, but the mineral has not yet been identified by the writer although reported in sericite schists of the Taconic Range by Dale (1899, p. 191). It may be that pyrophyllite, like paragonite, occurs as a concealed component in the white mica, a suggestion that is at least plausible, crystal-structurally.

Metamorphism of calcareous sediments

Siliceous calcite limestones, such as those in the Waits River formation appear to be relatively insensitive to the grade changes of ordinary regional metamorphism. With a small amount of argillaceous impurity, however, as in some of the Waits River beds, zoisite, and at slightly higher grades grossularitic garnet appear in impure limestones interstratified with schists containing almandite and, in some instances, staurolite and kyanite.

Siliceous dolomitic rocks are considerably more grade-sensitive, with the appearance of phlogopite in the chlorite zone, tremolite or actinolite in the garnet zone, and diopside in the staurolite-kyanite zone. In the presence of alumina clinozoisite or zoisite also appears in about the garnet zone. It is clear from the sequence; phlogopite-tremolite-diopside, that the magnesian component of the carbonates is the first to break down. In most of the Ca-Mg silicate rocks of the Chester dome area the remaining carbonate is almost entirely calcite as a result of this "de-dolomitization" process.

Metamorphism of igneous rocks

The metamorphism of felsic igneous rock types, like that of the low-alumina arenaceous sediments which they resemble compositionally, is relatively insensitive to changes in grade, a typical assemblage in any part of the area being: quartz-microcline-albite-biotite-muscovite, with either epidote or clinozoisite. More mafic types, however, such as the greenstones, are quite grade-sensitive. The typical mineral assemblage in the chlorite zone is: chlorite-epidote-albite-calcite. Ankerite may be present in addition to calcite, in some instances replacing it entirely. Quartz is generally present, and many types contain biotite or sphene in significant amounts. Actinolite appears about contemporaneously with garnet in the arenaceous rocks, apparently forming at the expense of both chlorite and carbonate. Epidote appears to increase in relative abundance at about the same time, apparently at the expense of calcite, and with the alumina probably derived from chlorite. In the staurolite-kyanite zone, chlorite is generally absent and carbonate much less abundant, with that remaining largely calcite. The amphibole in the staurolite-kyanite zone is characteristically of a strongly pleochroic, probably aluminous variety (Foslie, 1945). An almandite garnet is present in some occurrences in small amounts.

The mineral assemblages most characteristic of ultramafic rocks are: serpentine, talc-carbonate and talc-actinolite. The first two appear to be stable throughout the area, but actinolite has been observed only in the garnet and staurolite-kyanite zones. If, as the writer suspects, the fibrous talc in some of the Chester dome talc-carbonates is pseudomorphous after anthophyllite, it may mean that the assemblage anthophyllite-carbonate should replace talc-carbonate in the staurolite-kyanite zone, but is not preserved because of the relative instability (Bowen and Tuttle, 1949) of anthophyllite. How much of the serpentinization and related alteration is primary or the result of later metamorphism is a moot point, but the fact that the ultramafics of southern Vermont, occurring in regions of relatively high grade metamorphism, are in general completely serpentinized with considerable talc-carbonate alteration, in contrast to those farther north in regions of lesser metamorphism, suggests that a large part of the alteration must be secondary.

Retrograde metamorphism

Retrograde features, already described in a preceding section, are characteristic of nearly all the Green Mountain basement rocks, and are interpreted by the writer as the effects of the Paleozoic metamorphism superimposed upon a relatively higher grade pre-Cambrian metamorphism.

The mineral alterations and cataclastic features in the basement rocks are excellent examples of diapthoresis and phyllonitization, respectively. The almost complete absence of such features in the younger rocks, on the other hand, seems to indicate that there has been little or no recrystallization or deformation following the main period of Paleozoic metamorphism. Exceptions to this generalization include rare chlorite rims about garnet and the supposed pseudomorphs of talc after anthophyllite. The latter, however, may be related to the relative instability of anthophyllite, and the former, inasmuch as the chlorite rims are not only rare but very thin, to reaction with interstitial water during cooling.

It is significant that the basement rocks of the Chester dome do not show retrograde features and are essentially similar in texture and mineral composition to the younger rocks of the same area, a factor contributing greatly to the uncertainties in locating the basement unconformity in the dome. The fact that the basement rocks of the Chester dome show no clear evidence of an earlier metamorphism may be attributed to the higher grade attained by the younger metamorphism in the dome area, equalling or surpassing that of the earlier metamorphism, and the completeness of the recrystallization as indicated by the coarseness of grain of the younger rocks.

Movement of material

In the opinion of the writer there has been no large-scale metasomatism in any part of the present map-area. The various lithologic types, except for a progressive loss of volatiles with increasing grade, appear to retain their identity of chemical composition throughout the area, despite the conspicuous changes in mineralogy. The common occurrence of albite in schists has been taken as evidence of soda metasomatism but the writer considers such a conclusion contrary to the known facts concerning the chemical composition of argillaceous sediments. Barth (1936) has shown that schists containing up to 11% albite have about the bulk chemical composition of normal shales. The fact that the feldspar is entirely albite is apparently a reflection of the generalization that if a rock has an alkali-alumina ratio such that a feldspar and an aluminous mica must both be present, the potash will tend to be concentrated in the micas and the soda in the feldspars.

Alumina-rich schists such as those containing kyanite and staurolite, should be particularly susceptible to alkali metasomatism, leading to the formation of first, aluminous micas and later, feldspars. The fact that in the present map-area, thin horizons of such rock have retained their chemical identity over long distances in the Chester dome, and vicinity, seems to the writer to be inconsistent with any regional influx of alkalis. The highly aluminous types significant in this respect are the Pinney Hollow, Star Hill and Gassetts schists.

There does, however, appear to be ample evidence of small-scale metasomatism, of which probably the best examples are the reaction zones observed between rocks of contrasting chemical composition. These include tremolite or actinolite zones between dolomite and quartzite, actinolite, biotite and epidote between dolomite and schist, and the biotite, chlorite and actinolite zones between ultramafics and siliceous country rocks. The dolomite reaction zones are rarely more than two or three inches thick, but those about the ultramafics in the Chester dome area are commonly two or three feet thick. Phillips and Hess (1936) describe the reaction between serpentine and country rock as metamorphic differentiation. Although in essential agreement with their conclusions, the writer suggests that "homogenization" might be a better term for the process.

Further convincing evidence of small-scale movement of material is the reflection of the surrounding lithology in veins and boudinage-fillings. Similar features have been noted by Read (1934) in the Shetland Islands, and Niggli (1948) in the Alps. It is significant, however, that although the same major chemical elements are generally present in the vein or boudinage-filling as in the country rock, their proportions tend to differ slightly, and also significant that minerals rich in volatiles such as carbon dioxide and water are relatively more abundant in the segregations. Many such occurrences in relatively aluminous schists tend to have a composition approaching that of a pegmatite, indicating that alkalis have been concentrated in the opening with respect to alumina. The idea that relatively mobile constituents such as alkalis and volatiles may be effectively leached from the country rock in zones of tension leads to interesting speculations, particularly with regard to regions of higher grade metamorphism where mobility should, in general, be greater.

Although not generally considered metasomatic, the movements of carbon dioxide and water are also significant. Most rocks are "decarbonated" and "dehydrated" as a result of progressive regional metamorphism, but the ultramafics constitute an important exception to this rule, at least in the lower metamorphic grades, taking up first, water, in the serpentinization process, and later, carbon dioxide, in the alteration of serpentine to talc-carbonate. It would appear that talc-carbonate is more stable than serpentine in the presence of carbon dioxide at sufficient partial pressure, and significant, therefore, that although serpentinization is complete, the talc-carbonate alteration is only peripheral or along shear zones. Water, apparently, is the more mobile constituent.

Time of metamorphism

Rotational effects in porphyroblasts constitute the clearest evidence that much of the metamorphism was probably contemporaneous with the deformation. The mineralogic similarity of boudinage-fillings and veins to the surrounding rocks seems to indicate that their filling was contemporaneous with the metamorphism, although the time-lapse between the formation of the opening and its filling is difficult to evaluate. The writer is of the opinion, however, that the deformation and metamorphism were contemporaneous and related features, although the pressure and temperature conditions prevailing at the time of deformation may have continued for some time after its cessation.

Temperature and pressure conditions

It is beyond the scope of this study to attempt any determination of the physico-chemical conditions controlling the various metamorphic grades, inasmuch as experimental evidence relating to the stability ranges of the key minerals is inadequate. Recent work by Bowen and Tuttle (1949), however, has shown that serpentine is unstable much above 500°C at any geologically probable pressures, breaking down to forsterite and enstatite above that temperature. The effect of iron, present in all natural serpentines, is to lower the temperature of breakdown. Inasmuch as the serpentine at Chester apparently did not break down under the conditions prevailing in the staurolite-kyanite zone, it is reasonable to assume that the temperatures did not exceed 500° C, at least for any great length of time.

Metamorphism at Intrusive Contacts

The only metamorphism in the area that can definitely be related to an intrusive contact is that in the immediate vicinity of the Ascutney stock (Daly, 1903). The effects, particularly upon the phyllites of the Northfield and Waits River formations, are quite striking, although limited to a zone generally not more than a few hundred yards wide along the contact. Apparently quartz is about the only major mineral constituent of the phyllites stable in the aureole. The almandite garnets are replaced by pseudomorphs of iron-rich biotite, spinel and cordierite, and the micas of the groundmass by cordierite, sillimanite, alkali feldspar, and, in some instances, corundum and spinel. At the immediate contact the rock is a dense, bluish hornfels with but a faint streakiness suggesting the original schistosity.

The other intrusive bodies of the area, the ultramafics, and the small pegmatitic and granitic masses, however, show no such evidence of extreme high-temperature contact effects. Where some of the small pegmatitic and granitic bodies are in contact with marbles, it is true that there is an abnormal abundance of silicate in the marble, sometimes for a distance of several feet from the contact, but it is significant that the silicates present are the same as those in surrounding areas, clearly related to the regional metamorphism. If high-temperature minerals such as wollastonite or forsterite ever existed at these contacts, they have presumably been destroyed by later metamorphism. The writer, therefore, attributes the lack of high-temperature contact effects about the older intrusive bodies largely to the fact that they were probably emplaced prior to the regional metamorphism, or at least early enough in the regional metamorphism, so that such effects have not been preserved.

MAJOR PROBLEMS

Source of the Taconic Klippe

The source of the Taconic klippe is one of the more baffling problems of the geology of western New England (Prindle and Knopf, 1932; Hawkes, 1941). Although an eastern source is indicated, no satisfactory root zone has yet been found (Hawkes, 1941).

The writer's contributions to the problem consists largely of the elimination of the areas mapped in detail as possible root zones, and the stratigraphic evidence that the Taconic sequence probably represents a more western facies than that of the Ludlow area. Although the evidence available seems to point toward the Green Mountain region as the source area, recent work by P. H. Osberg and W. M. Cady at the northern end of the anticlinorium has failed to disclose the existence of any large, through-going fault zone, as has the writer's reconnaissance in the Wallingford quadrangle. A hypothesis generally credited to W. S. White, however, is that the klippe is a rootless mass possibly representing a gigantic landslip off the crest of the rising Green Mountain anticlinorium and the dome-area to the east. In the light of the present evidence this hypothesis seems as plausible as any and is that which the writer is inclined to favor. It is interesting that a similar origin has recently been suggested for some of the Alpine decken (H. Cloos, 1949, lecture at Massachusetts Institute of Technology).

Origin of the Chester Dome

The problem of the origin of the Chester dome is perhaps fully as baffling as that of the source of the Taconic klippe. The evidence seems to be incontrovertible that a mass of ancient crystalline rock, mantled by younger strata, has moved almost vertically upward for a considerable distance, with attendant intense stretching and metamorphism of the mantle-rock. The causes of such a movement, however, are far from clear.

One plausible suggestion, compatible with most conventional tectonic theory, is that the dome-rock has been squeezed upward, like toothpaste from a tube, as the result of lateral movements at depth. Another suggestion, emphasizing the structural resemblance to a salt dome, is that the upward movement may be related to a density difference between the dome-gneisses and the surrounding schists. The writer determined the density of seven specimens of the coarse phase of the Reading gneiss from the center of the dome. The values ranged from 2.66 to 2.73, averaging about 2.69. Density determinations by R. J. Bean (personal communication) upon typical specimens of the younger formations from the regions immediately north of the dome, show that the schists constituting the bulk of the younger rock average somewhat higher, and that 2.80 is probably a representative figure. Gravimetric measurements by Bean, furthermore, show gravity lows in the vicinity of the two small "buried" domes (plate I) northeast of the Chester dome, indicating that the difference is real. Whether the difference is of sufficient magnitude to cause the doming, however, is by no means certain. The key unknowns are probably the plasticity of the rocks at the time of deformation and the length of time involved. Many of the structural features of the dome appear to suggest high fluidity but "appearances" cannot, unfortunately, be quantitatively evaluated.

The magmatic origin currently in favor (C. A. Chapman, 1942) for the Oliverian domes of western New Hampshire does not appear to be applicable in any way to the Chester dome, despite many obvious similarities. It is interesting that in the early work of C. H. Hitchcock (1877, 1890, 1908) the Oliverian dome-gneisses were interpreted as older than the surrounding rocks, although Hitchcock later (1912) reversed his earlier interpretation and concluded that they were igneous intrusives. The work of B. K. Emerson in Massachusetts shows a similar evolution of ideas, the dome gneisses first interpreted (1892, 1898) as an older "conglomerate-gneiss" and later as the intrusive "Monson granodiorite" (1917). Most of the domes of Monson granodiorite appear to be a continuation of the Oliverian domes of New Hampshire. The dome at Shelburne Falls, however, is considerably west of the others and about on strike with the Chester dome. It has recently been re-studied by Balk (1946), whose interpretation of the gneisses as representing an older basement complex is in agreement with that of the Chester dome.

A recent review of the gneiss-dome problem by Eskola (1949) shows that the interpretation of gneiss-domes in other areas, notably the Finnish Archaean, has gone through similar cycles. It is rather surprising that geologists, at least English and American, have been slow to recognize the gneiss-dome problem as such, papers such as those of C. H. Hitchcock (1890), Quirke and Lacy (1941) and the recent review by Eskola, being relatively rare.

The interpretation favored by Eskola is that the domes are, for the most part, granitic rocks clearly older than the mantle-rocks, and that their later doming has been aided by the addition of new granitic material with attendant migmatization and granitization, and in some instances actual rheomorphism and flowage, as an intrusive mass, of the dome-gneiss. There is no clear evidence, however, in the Chester dome, that the basement gneisses ever were intrusive granites. The core rocks include many obvious sedimentary types and the gneisses do not show intrusive relationships to these. The granites and pegmatites that do occur in the basement complex are but a small fraction of the total rock-mass in the areas studied by the writer. A rather disturbing suggestion is that the basement gneisses may themselves be old dome gneisses! The need for a thorough study of the Green Mountain basement structures and their relationship to the Chester dome is apparent.

The writer is also doubtful as to the causal significance of the metamorphism and introduction of younger granitic material. In the Chester dome the younger granitic rocks appear to be somewhat older than the metamorphism, and therefore not necessarily related to the doming. It is possible that some of the similar, but larger, granitic bodies in the regions east of Barre may be related to buried gneiss-domes but such a relationship has yet to be demonstrated. There is little evidence, furthermore, that the formation of the Chester dome was accompanied by any large-scale metasomatic activity. Regardless of the final outcome, however, it is evident that the solution of the gneiss-dome problem will be of tremendous petrogenetic and tectonic significance.

BIBLIOGRAPHY

Bain, G. W. (1936) Serpentinization of Vermont ultrabasics. Geol. Soc. Amer. Bull. Vol. 47, p. 1961-1980.

(1938) Correlatives of the Grenville Series. Geol. Soc. Amer. Bull. Vol 44, p. 1807-1827.

(1942) Vermont talc and asbestos deposits. Ore deposits as related to structural Features. Newhouse, ed. p. 255-258.

Balk, Robert (1936) Devitrified felsite dikes from Ascutney Mts., Vt. Am. Mineralogist Vol. 21, p. 516-522.

(1936) Structural and petrologic studies in Dutchess County, N.Y. Part I. Geologic structure of sedimentary rocks. Geol. Soc. Amer. Bull. Vol. 47, p. 685-774.

(1942) The Pelham gneiss dome, Massachusetts (abstract). Am. Geophys. Union Trans. Pt II P. 343-344.

(1946) Gneiss dome at Shelburne Falls, Mass. Geol. Soc. Amer. Bull. Vol. 57, p. 125-160.

Barth, T. F. W. (1936) Structural and petrologic studies in Dutchess County, N. Y. Part II. Petrology and metamorphism of the Paleozoic rocks. Geol. Soc. Amer. Bull. Vol. 47, p. 775-850.

Billings, M. P. (1937) Regional metamorphism of the Littleton-Moosilauke area, N. H. Geol. Soc. Amer. Bull. Vol. 48, p. 463-566.

Bowen, N. L. and Tuttle, O. F. (1949) The system $MgO - SiO_2 - H_2O$. Geol. Soc. America Bull. Vol. 60, p. 434-460.

Chapman, R. W. and Chapman, C. A. (1940) Cauldron subsidence at Ascutney Mountain, Vermont. Geol. Soc. America Bull. Vol. 51, P. 191-212.

Chapman, C. A. (1942) Intrusive domes of the Claremont-Newport Area, N.H. Geol. Soc. America Bull. Vol. 53, P. 889-916.

Clark, T. H. and Fairbairn, H. W. (1936) The Bolton Igneous Group of Southern Quebec. Royal Soc. Canada Trans., 3rd Ser., Sect. IV, Vol. 30, p. 13-18

Clark, T. H. (1936) A Lower Cambrian series from southern Quebec. Royal Canadian Inst. Trans. Vol. 21, Pt. 1 p. 135-151.

Clarke, F. W. (1924) The data of geochemistry. 5th edition U.S. Geol. Survey, Bull. #770, p. 841.

Cloos, E. (1946) Lineation: a critical review and annotated bibliography. Geol. Soc. America Mem. 18, p. 122.

(1947) Boudinage Am. Geophys. Union Trans. Vol. 28, p. 626-632.

Cooke, H. C., (1937) Thetford, Disraeli and eastern half of Warwick map-areas, Quebec, with chapters on the Benceville, St. Francis and Lake Aylmer series by Thomas Henry Clark. Canada Geol. Survey Mem. 211, p. 160

Currier, L. W. (1934) <u>Notes on staurolite and associated minerals from schists at Gassetts, Vt.</u> Am. Mineralogist Vol. 19, p. 335-339.

Currier, L. W. and Johns, R. H. (1941) <u>Ordovician stratigraphy of central Vermont</u>. Geol. Soc. America Bull. Vol. 52, P. 1487-1512.

Dale, T. N. (1894) <u>On the structure of the ridge between the Taconic and Green Mountain ranges in Vermont</u>. U. S. Geol. Survey, 14th Ann. Rept. Pt. 2, p. 525-549.

(1896) <u>Structural details in the Green Mtn. region</u> and in eastern New York. U. S. Geol. Survey, 16th Ann. Rept. Pt. 1, p. 543-570.

(1899) <u>The slate belt of eastern New York and western Vermont</u>. U. S. Geol. Survey 19th Ann. Rept. Pt. 3, p. 159-306.

(1902) <u>Structural details in the Green Mountain</u> region and in eastern New York (second paper). U. S. Geol. Survey Bull. 195, p. 10-13.

(1910) <u>The Cambrian conglomerate of Ripton in Vermont</u>. Am. Jour. Sci. 4th Ser. Vol. 30, p. 267-270.

(1912) <u>The commercial marbles of western Vermont</u>. U. S. Geol. Survey Bull. p. 170.

(1915) <u>The calcite marble and dolomite of eastern Vermont</u>. U. S. Geol. Survey, Bull. 589, p. 67.

Dale, T. N. (1916) <u>The Algonkian-Cambrian boundary east of the Green Mtn. axis in Vermont</u>. Am. Jour. Sci. 4th ser. Vol. 42, p. 120-124.

Daly, R. A. (1903) <u>Geology of Ascutney Mountain, Vermont</u> U. S. Geol. Survey Bull. 209, p. 122.

Doll, C. G. (1943) <u>A Paleozoic revision in Vermont.</u> Am. Jour. Sci. Vol. 241, p. 57-64.

(1943) <u>A brachiopod from mile schist, South Strafford, Vt.</u> Am. Jour. Sci. Vol. 341, p. 676-679.

(1944) <u>A preliminary report on the geology of the Strafford quadrangle, Vermont</u>. Vermont State Geologist 24th Rept., p. 14-28.

Eggleston, J. W. (1918) <u>Eruptive rocks at Cuttingsville, Vermont</u>. Am. Jour. Sci. 4th ser. Vol. 45, p. 377-410.

Emerson, B. K. (1892) <u>Outlines on geology of the Green Mountain region in Massachusetts</u>; Hawley sheet discriptive text. U. S. Geol. Survey, Geol. Atlas, Hawley sheet, p. 3.

. . . (1898) <u>Geology of old Hampshire County, Mass</u>. U. S. Geol. Survey Mon. 29, p. 790.

(1917) <u>Geology of Massachusetts and Rhode Island.</u> U. S. Geol. Survey Bull 597, p. 289

Eskola, P. E. (1920) <u>The mineral facies of rocks.</u> Norsk. Geol. Tidsskr. Vol. 6, p. 143-194.

(1949) <u>The problem of mantled gneiss domes</u>. Quart. Jour. Geol. Soc. London, Vol. 104 Pt. 4, p. 461-476

Fairbairn, H. W. (1932) Structure and metamorphism of Broome County, Quebec. Harvard University, doctorate thesis, p. 164.

(1935) Notes on the mechanics of rock foliation. Jour. Geol. Vol. 43, p. 591-608.

(1936) Elongation in deformed rocks. Jour. Geol. vol. 44, p. 670-680.

Foslie, Steinar (1945). Hastingites, and amphiboles from the epidote-amphibolite facies. Norsk. Geol. Tidsskr. Vol. 25, p. 74-98.

Fowler, P. (1949) Stratigraphy and structure of the Castleton area, Vt. Harvard University, doctorate thesis, p. 102.

Foye, W. G. (1919) A report on the geological work within the Rochester, Vermont, quadrangle. Vermont State Geologist 11th Rept. p. 76-98.

Foyles, E. J. (1928) Rock-correlation studies in west-central Vermont. Vermont State Geologist, 16th Rept. p. 281-284.

(1931) The geology of East Mountain, Mendon, Vt. Vermont State Geologist 17th Rept. p. 238-251.

(1931) Compressed mild resembling graptolites. Vermont State Geologist 17th Rept. p. 252.

(1933) Metamorphism near Rutland, Vt. Vermont State Geologist 18th Rept. p. 363-382.

Gillson, J. L. (1927) Origin of the Vermont talc deposits, with a discussion on the formation of talc in general. Econ. Geology Vol. 22, p. 246-287.

Gregory, H. E. and Robinson, H. H. (1907) Preliminary geological map of Connecticut. Connecticut State Geol. & Nat. Hist. Survey Bull. 7, p. 39.

Grubenmann, V. and Niggli, P. (1924) Die Gesteinsmetamorphose Berlin, Verlag von Gebruder Borntraeger, p. 539.

Hadley, J. B. (1942) Stratigraphy, structure and petrology of Mt. Cube area, N.H. Geol. Soc. America Bull. Vol. 53, p. 113-176.

Harker, A. (1932) Metamorphism. London, Methuen & Co., Ltd. p. 362.

Hawkes, H. E., Jr. (1941) Roots of the Taconic Fault in west-central Vermont. Geol. Soc. America Bull. Vol. 52, p.649-660.

Heim, A. (1878). Mechanismus der Gebirgsbildung, Vol. II. Basel, Benno Schwabe, p. 246.

Hess, H. H. (1933) The problem of serpentinization and the origin of certain chrysotile asbestos, talc, and soap-stone deposits. Econ. Geology Vol. 28, p. 634-657.

Hitchcock, C. H. (1877) Geology of New Hampshire, Part II. Stratigraphical geology. Concord, N. H., p. 684.

(1890) Significance of oval granitoid areas in the lower Laurentian. (abstract with discussion by G. H. Williams). Geol. Soc. America Bull. Vol. 1., p. 557-558.

Hitchcock, C. H. (1908) Geology of the Hanover, N.H. quadrangle. Vermont State Geologist 6th Rept. p. 139-186

(1912) The Strafford quadrangle. Vermont State Geologist 8th Rept. p. 100-145.

Hitchcock, E., Hitchcock, E. J., Hager, A. D., Hitchcock, C. H. (1861) Report on the Geology of Vermont. Claremont, N.H. p. 982.

Kaiser, E. P. (1945) Northern end of the Taconic thrust-sheet in western Vermont. Geol. Soc. America Bull. Vol. 56, p. 1079-1098.

Keith, A. (1932) Stratigraphy and structure of north-western Vermont. Washington Acad. Sci. Jour. Vol. 22, p. 357-379, 393-406.

Kruger, F. C. (1946) Structure and metamorphism of the Bellows Falls quadrangle of N.H. and Vt. Geol. Soc. America Bull. Vol. 57, p. 161-206.

Leith, C. K. (1905) Rock cleavage. U. S. Geol. Survey Bull. 239, p. 216.

Mackay, B. R. (1921) Beauceville map-area, Quebec. Canada Geol. Survey Mem. 127, p. 105.

Maynard, J. E. (1934) The petrographic re-examination of quartz-bearing plutonites from Vermont. Jour. Geol. Vol. 42, p. 146-162.

Mead, W. J. (1940) Folding, rock flowage and foliate structures. Jour. Geol. Vol. 48, p. 1007-1021.

Morre, G. E. (1949) Structure and metamorphism of the Keene-Brattleboro area, New Hampshire-Vermont. Geol. Soc. America Bull. Vol. 60, p. 1613-1670.

Niggli, P. (1948) Some aspects of the Geol., petrol, & min. of Switzerland. Washington Acad. Sci. Jour. Vol. 38, p. 289-306.

Perry, E. L. (1928) The geology of Bridgewater and Plymouth townships, Vermont. Vermont State Geologist 16th Rept. p. 1-64.

Phillips, A. H. and Hess, H. H. (1936) Metamorphic differentiation at contacts between serpentinite and siliceous country rocks. Am. Mineralogist Vol. 21, p. 333-362.

Phillips, F. C. (1947) Lineation in the Northwest Highlands of Scotland. Geol. Magazine Vol. 84, p. 58-59

Prindle, L. M. and Knopf, E. B. (1932) Geology of the Taconic quadrangle. Am. Jour. Sci. 5th ser. Vol. 29, p. 257-302.

Pumpelly, R., Wolff, J. E., Dale, T. N. (1894) Geology of the Green Mountains in Massachusetts. U. S. Geol. Survey Mon. 23, p. 206.

Quirke, T. T., Lacy, W. C. (1941) Deep-zone dome and basin structures. Journ. Geol. Vol 49, p. 598-609.

Read, H. H. (1934) On the segregation of quartz-chlorite-pyrite masses in Shetland igneous rocks during dislocation metamorphism with a note on the occurrence of boudinage structure. Liverpool Geol. Soc. Proc. Vol. 16, p. 134-138.

Richardson, C. H. (1902) The terranes of Orange County, Vermont. Vermont State Geologist, 3rd Rept. p. 61-101.

Richardson, C. H. and Camp, S. H. (1919) The terranes of Northfield, Vermont. Vermont State Geologist, 11th Rept. p. 99-119.

Richardson, C. H. (1927) The geology and petrography of Barnard, Pomfret, and Woodstock, Vt. Vermont State Geologists 15th Rept. p. 127-159.

Richardson, C. H. (1928) The geology and petrography of Reading, Cavendish, Baltimore and Chester, Vt. Vermont State Geologist 16th Rept. p. 208-248.

(1931) The areal and structural geology of Springfield, Vt. Vermont State Geologist 17th Rept. p. 193-212.

(1931) The geology and petrography of Grafton and Rockingham, Vt. Vermont State Geologist 17th Rept. p. 213-237.

Richardson, C. H. and Maynard, J. E. (1938) Geology of Vernon, Guilford and Halifax townships, Vt. Vermont State Geologist, 21st Rept. p. 74-83.

Ruedemann, R. (1930) Geology of the capital district. New York State Mus. Bull. 285, p. 218.

(1947) Graptolites of North America. Geol. Soc. America Mem. 19, p. 652.

Tolman, C. (1936) Lake Etchemin map-area, Quebec. Canada, Canada, Geol. Survey Mem. 199, p. 20.

Turner, F. J. (1948) Evolution of the metamorphic rocks. Geol. Soc. America Mem. 30, p. 342.

Van Hise, C. R. (1896) Principles of North American pre-Cambrian geology. U. S. Geol. Survey 16th Ann. Rept. p. 571-874.

Walcott, C. D. (1888) The Taconic system of Emmons, and the use of the name Taconic in geologic nomenclature. Am. Jour. Sci. 3rd ser., Vol. 35, p. 292-242, 307-327, 394-401.

Weiss, J. (1949) Wissahickon Schist at Philadelphia, Pennsylvania. Geol. Soc. America Bull. Vol. 60, p. 1689-1726.

White, W. S. (1946) Structural geology of the Vermont portion of the Woodsville quadrangle. Harvard University, doctorate thesis, p. 166.

(1949) Cleavage in east-central Vermont. Am. Geophys. Union Trans. Vol. 30, p. 587-594.

Whittle, C. L. (1892) An ottrelite-bearing phase of a metamorphic conglomerate in the Green Mts. Am. Jour. Sci. 3rd ser. Vol. 44.

(1893) Some dynamic and metasomatic phenomena in a metamorphic conglomerate in the Green Mts. Geol. Soc. America Bull. Vol. 4, p. 147-166.

Whittle, C. L. (1894) The general structure of the main axis of the Green Mountains. Am. Jour. Sci. 3rd ser. Vol. 47, p. 347-355.

(1894) The occurrence of Algonkian rocks in Vermont and the evidence for their subdivision. Jour. Geol. Vol. 2, p. 396-429.

Williams, C. R. and Billings, M. P. (1938) Petrology and structure of the Franconia quadrangle, N. H. Geol. Soc. America Bull. Vol. 49, p. 1011-1044.

Wolff, J. E. (1891) On the Lower Cambrian age of the Stockbridge limestone. Geol. Soc. America Bull. Vol. 2, p. 331-337.

Biographical Data

Name: James Burleigh Thompson, Jr.

Born: Calais, Maine, November 20, 1921.

Institutions attended:

Public Schools, Fort Lee, New Jersey, 1927-1938

Dartmouth College, 1938-1942, A.B. 1942

Massachusetts Institute of Technology, Jan.-Sept. 1943 and March, 1946 to Jan. 1950.

Degrees and honors:

A.B., cum laude, in geology, Dartmouth College, 1942

Military service:

U.S. Army Air Forces, December, 1942 to February, 1946 (meteorology)

Professional experience:

Instructor in Geology, Dartmouth College, summer and fall terms, 1942.

Research Assistant in Geology, M.I.T., 1946-47

Instructor in Geology, M.I.T., 1947-49.

Instructor in Geology, Harvard University, July, 1949 to present.

Scientific and professional societies:

Sigma Xi

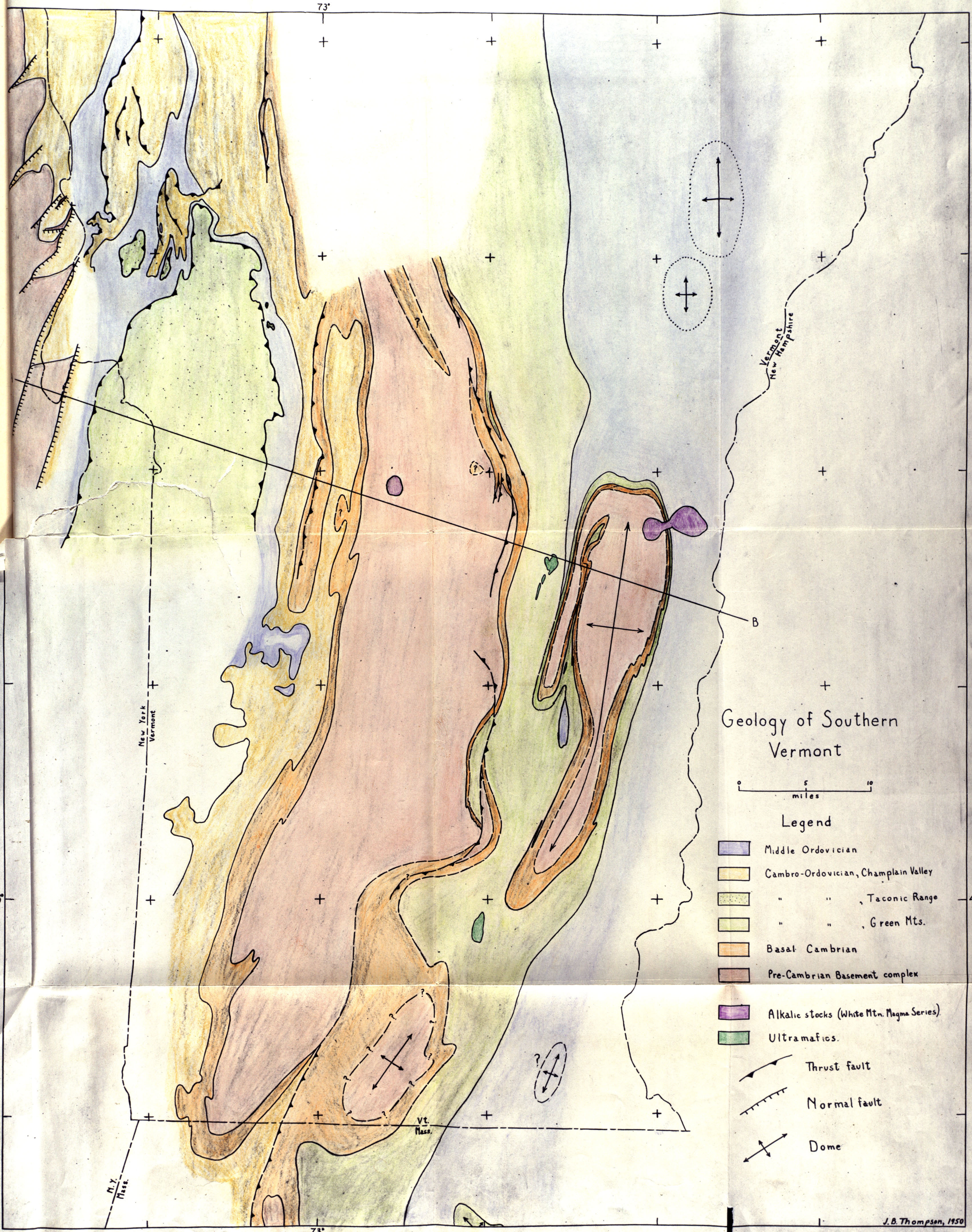
American Association for the Advancement of Science.

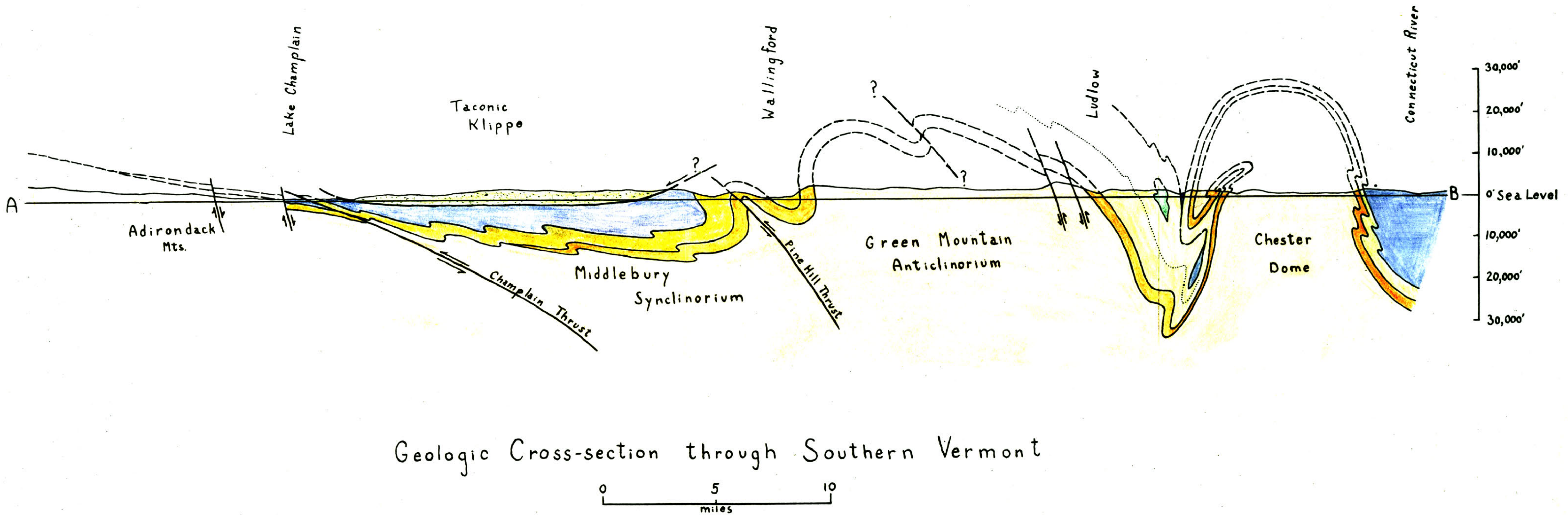
Geological Society of America (member)

Publications:

Role of aluminum in the rock-forming silicates. (abstract) --Geol. Soc. America Bulletin, vol. 58, p. 1232,1947

Plate I.

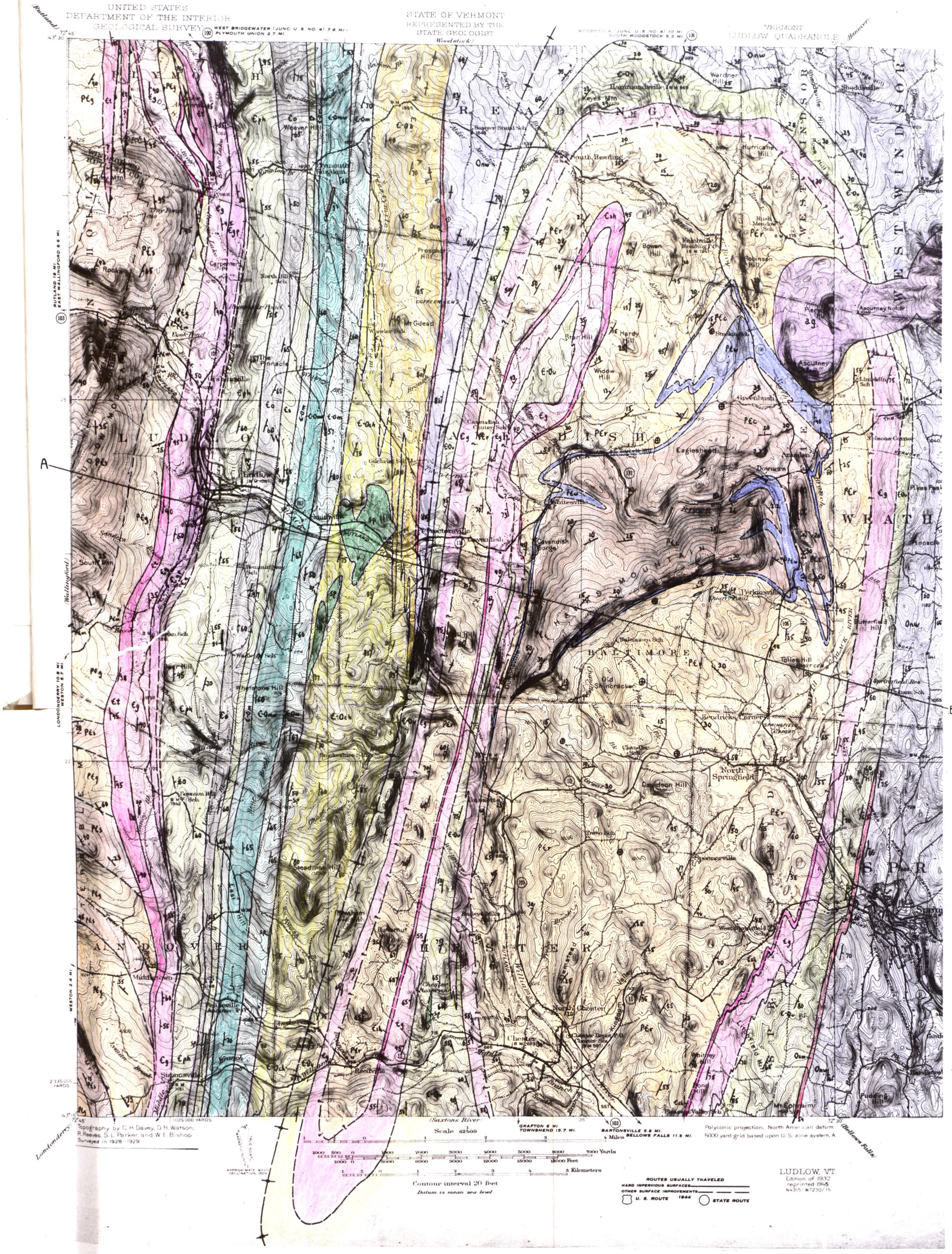

73°
44°
43°
43°
73°
B
Vermont
New Hampshire
New York
Vermont
Vt.
Mass.
N.Y.
Mass.
Geology of Southern Vermont
0 5 10
miles
Legend
Middle Ordovician
Cambro-Ordovician, Champlain Valley
" " , Taconic Range
" " , Green Mts.
Basal Cambrian
Pre-Cambrian Basement complex
Alkalic stocks (White Mtn. Magma Series)
Ultramafics.
Thrust fault
Normal fault
Dome
J.B. Thompson, 1950



Geologic Cross-section through Southern Vermont

No vertical exaggeration
(Legend same as map.)

J.B. Thompson, 1950



GEOLOGIC MAP OF THE LUDLOW AREA

LEGEND

Stratified rocks, eastern limb of Green Mountain anticlinorium:

Middle Ordovician
- Onw — Northfield and Waits River formations.
- Osm — Shaw Mountain formation.

--------Unconformity--------

Cambrian or Lower Ordovician
- Є-Och / Є-Ob — Cram Hill formation (C-Och); and Barnard gneiss (C-Ob).
- Є-Om / Є-Omw — Moretown formation (C-Om); with Whetstone Hill member (C-Omw).

Probably Lower Cambrian
- Єs — Stowe formation.
- Єo — Ottauquechee formation.
- Єph — Pinney Hollow formation.
- Єgp / Єg — Grahamville formation (Cg); with Plymouth member (Cgp).
- Єt — Tyson formation.

--------Angular Unconformity--------

Pre-Cambrian
- PЄs — Schists and quartzites.
- PЄm — Marbles and Ca-Mg silicate rocks.
- PЄg — Gneisses.

Pre-Shaw Mountain stratified rocks in Chester dome:

Cambrian or Lower Ordovician
- Є-Ou — Pinney Hollow fm. through Cram Hill fm., undivided.

Probably Lower Cambrian
- Єg — Grahamville formation.
- Єsh — Star Hill formation.

----?---Angular Unconformity---?----

Probably Pre-Cambrian
- PЄc — Cavendish and Gassetts schists.
- PЄw — Whitesville marble.
- PЄr — Reading gneiss.

Unstratified rocks:

Post-metamorphism
- ag — Ascutney gabbro and syenite.

Pre-metamorphism
- qd — Quartz-diorite.
- gs — Greenstone.
- sp / sp — Serpentine and talc-carbonate rock.

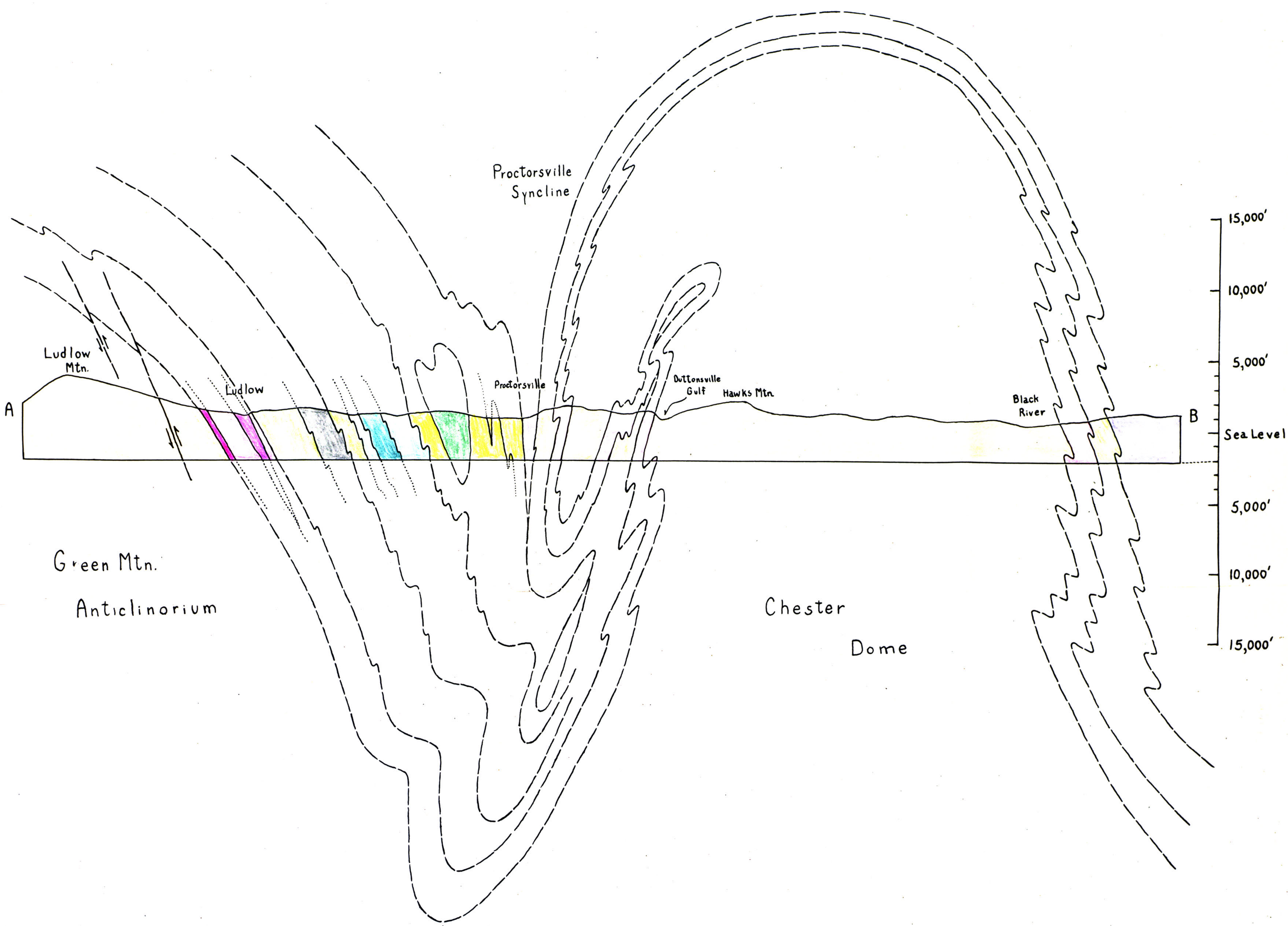
Contacts:

- Good control.
- Fair control.
- Poor control.

Structural features:

- Thrust fault, sawteeth on overthrust side.
- Axis of Proctorsville syncline.
- 20 — Strike and dip of compositional banding.
- Strike of vertical compositional banding.
- ⊕ — Compositional banding horizontal.

Geology by J. B. Thompson, 1947-49, assisted by P. H. Chang and J. Gieling, 1947; R. J. Bean and K. N. Das, 1948; and A. L. Albee, J. W. Skehan and M. C. Wittels, 1949.

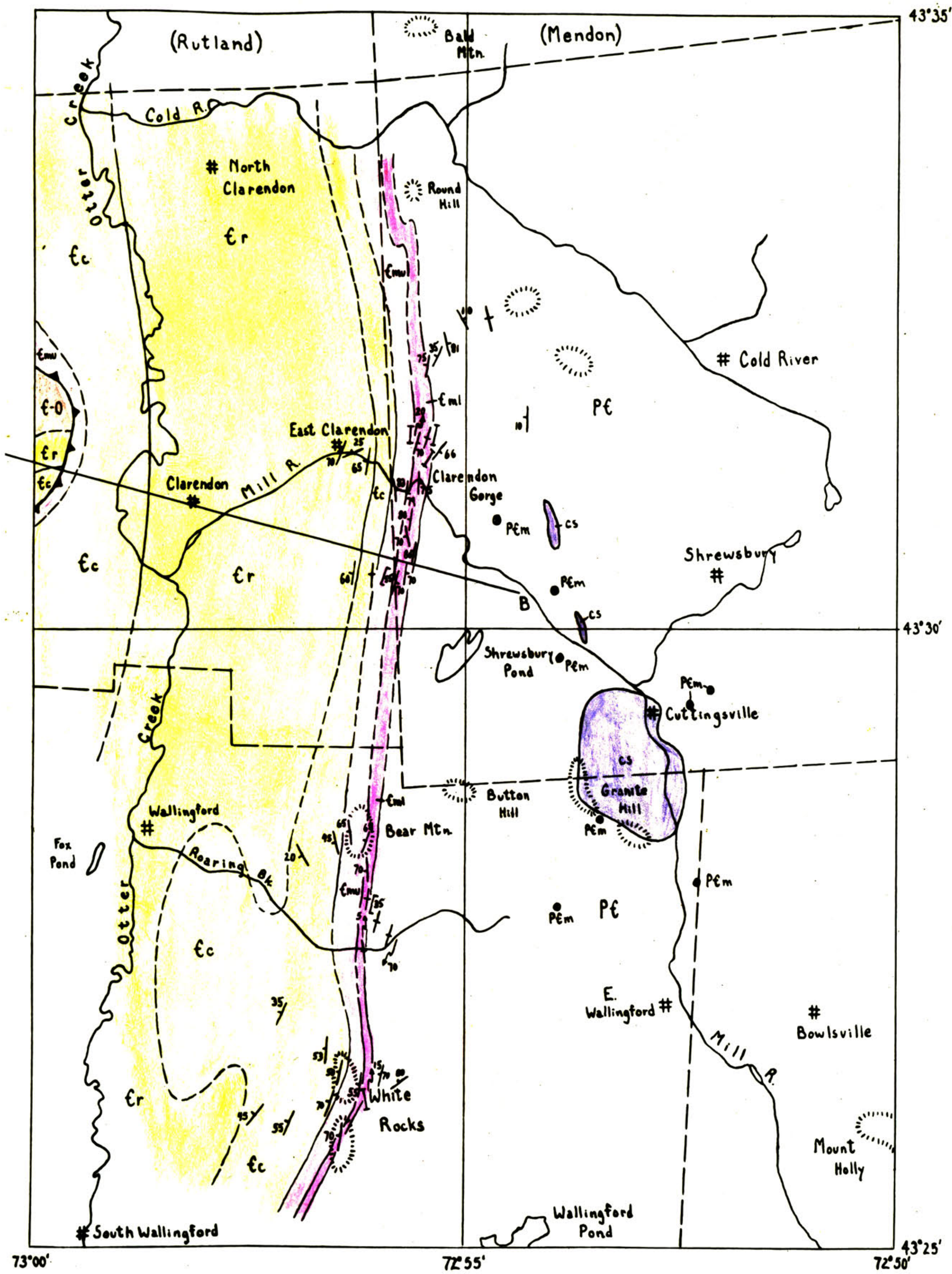


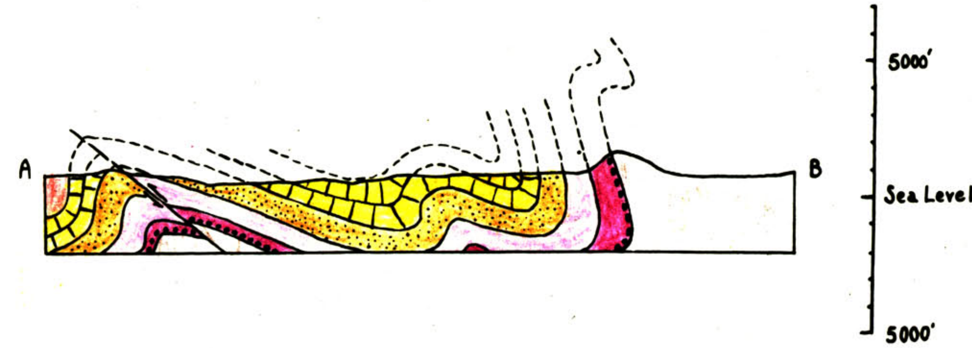
Geologic Cross-section of the Chester Dome

0 1 2 3 4
miles

No vertical exaggeration.
Legend same as map.

J.B. Thompson, 1950





Cross-section along line A-B.
(Same scale as map.)

Legend:

Lithology

Late Paleozoic	cs	Cuttingsville syenite and associated rocks.
Cambro-Ordovician	Є-O	Sedimentary formations younger than Rutland dol.
Lower Cambrian	Єr	Rutland dolomite.
	Єc	Cheshire quartzite.
	Єmu	Upper Mendon Series.
	Єml	Lower Mendon Series.
		(Angular Unconformity)
Pre-Cambrian	PЄ	Undifferentiated schists and gneisses.
	PЄm	Marbles and Ca-Mg silicate rocks.

Structure

30 Strike and dip of compositional banding.

75 Strike and dip of slip-cleavage.

20 Plunge and shear-sense of minor fold.

Lithologic boundary.

Thrust fault.

Geology based in part upon maps by J.E. Wolff (1891); J.W. Eggleston (1918); and Phillip Fowler (1949).

Geologic Map of the Wallingford-Clarendon Area

0 1 2
miles

J.B. Thompson, 1950

Plate III.

TECTONIC MAP OF THE LUDLOW AREA

LEGEND:

- Strike and dip of compositional banding.
- Strike of vertical compositional banding.
- Compositional banding horizontal.
- Strike and dip of foliation.
- Strike of vertical foliation.
- Strike and dip of slip-cleavage.
- Strike of vertical slip-cleavage.
- Plunge of minor fold.
- Plunge and shear-sense of minor fold.
- Mineral-lineation.
- Boudinage rupture-line.
- Long axes of deformed pebbles.
- Long axes of deformed orbicules.
- Axis of Proctorsville syncline.
- Axis of Butternut Hill fold.
- Axis of Chester dome.
- Thrust fault, sawteeth on overthrust side, dashed where position doubtful.
- Major lithologic boundaries.

J. B. Thompson, 1950